THE AMERICAN MINERALOGIST
JOURNAL OF THE MINERALOGICAL SOCIETY OF AMERICA

Vol. 23 APRIL, 1938 No. 4

X-RAY IMAGE METHOD OF CHEMICAL ANALYSIS

L. v. HÁMOS, *Stockholm, Sweden.*

The earlier methods of x-ray spectroscopy have recently been supplemented by methods in which curved crystal surfaces are used to produce the spectra. Most of these methods (M. de Broglie and F. A. Lindemann; H. Johann;[1] Y. Cauchois[2]) give line spectra of the usual appearance. The advantage of these so called "focusing" spectrographs is the increased sensitivity.

As the author pointed out,[3] curved crystals can be used also for producing a new kind of x-ray spectrum consisting of a series of two dimensional images instead of the spectral lines. Each of these images is formed by x-rays of a definite wave length. They may, therefore, be called "spectral images." In the following the general features of x-ray image spectroscopy will be given.

I. The Production of True X-Ray Images

For producing true images in ordinary optics either a lens or a mirror is needed. As the refraction of x-rays is very small, it is hardly possible to build a lens for x-rays. On the other hand, the diffraction of x-rays by crystals, which according to W. L. Bragg can be interpreted as reflection of these rays on crystal planes, permits the realization of an x-ray mirror producing true images.

If an x-radiation of wave length λ is incident on a crystal surface, a selective reflection takes place at the angle ϕ given by the Bragg formula:

$$n\lambda = 2d \cdot \sin\phi.$$

In this equation d means the grating constant belonging to the reflecting crystallographic plane and n is the order of reflection.

We can now compare in Fig. 1 the path of the rays reflected by a plane crystal surface with the path of rays reflected by a cylindrical concave crystal mirror.

In Fig. 1a the source O of the x-rays emits radiation of wave length λ. The plane crystal K reflects the rays with the glancing angle ϕ. The divergent reflected bundle of x-rays forms a part of a cone. Its intersec-

[1] Johann, H., *Zeits. Phys.*, vol. **69**, p. 185, 1931.
[2] Cauchois, Y., *Jour. Phys. et Radium*, vol. **7**, III, p. 320, 1932.
[3] Hámos, L. v., *Nature*, vol. **134**, p. 181, 1934.

tion with the plane F, parallel to the crystal K and containing the point O, is a segment of a circle C.

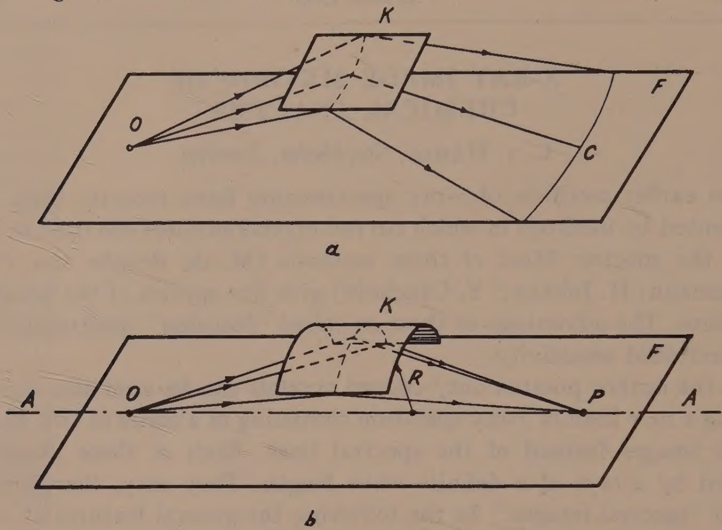


FIG. 1. Reflection of x-rays by (a) a plane, and (b) by a cylindrical crystal mirror.

In Fig. 1b the reflection takes place on a cylindrical crystal surface. The path of the reflected x-rays depends on the structure of the crystal. In order to get a convergent reflected bundle the structure of each surface element of the concave side of the crystal has to be orientated tangentially to the cylindrical surface. If this condition is satisfied, the rays having their origin in the point O (lying on the cylinder axis A–A) will meet again after reflection on the concave side of the cylindrical crystal K in the point P (lying also on the axis A–A). P is a true x-ray image of O.

It can easily be shown that even from points lying in the vicinity of the axis A–A true x-ray image-points will be formed. Figure 2 indicates the relative position of a two-dimensional x-ray source O, of a concave crystal mirror K, and of two different x-ray images B_1 and B_2 produced

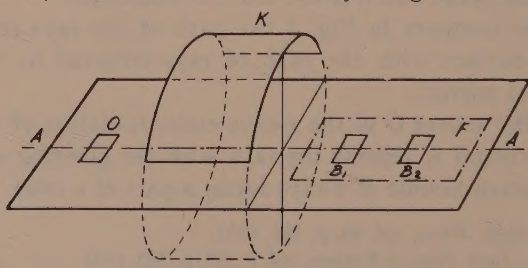


FIG. 2. Relative position of an x-ray source and its image spectrum.

by K. Each of these images corresponds to a definite wave length component of the radiation emitted by O.

If the source of x-rays is homogeneous, all parts of it emit radiation of identical spectral composition. The images of such a source will be of the same size as the source. In this special case there is consequently no essential difference between a line spectrum of the usual form and our image spectrum.

Let us now consider in Fig. 3 a schematical example of an x-ray source which is not homogeneous. We suppose that there are three different phases in the sample: A, B and $A+B$. The wave length emitted by A may be called λ_A, and the wave length emitted by B, λ_B The phase $A+B$ will emit both λ_A and λ_B. The image spectrum corresponding to the sample is also indicated in Fig. 3. In the spectral image corresponding to the wave length λ_A only the parts of the sample containing the element A appear. In the same manner the λ_B-image represents the parts of the sample containing the element B.

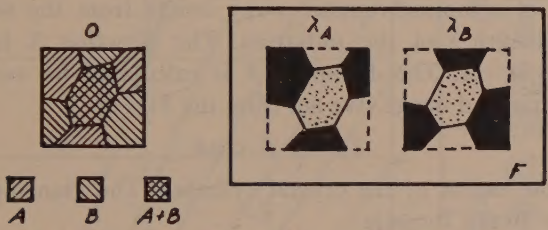


FIG. 3. Schematical example for an image spectrum.

We see from this simple example that x-ray images formed by a crystal mirror can be of different appearance, which is a new feature of x-ray image spectroscopy. We shall see later on that new possibilities of chemical analysis arise from this property of image spectra.

II. THE X-RAY IMAGE SPECTROGRAPH

An x-ray spectrograph operating with a cylindrical crystal mirror is schematically indicated in Fig. 4. Q is an x-ray tube producing the pri-

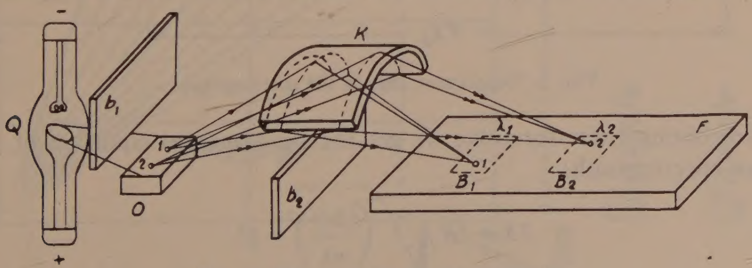


FIG. 4. The x-ray image spectrograph.

mary x-rays. O is the sample to be analyzed, having a plane surface. F is a photographic plate (or film) while b_1 and b_2 are diaphragms limiting the x-ray bundles.

The apparatus acts as follows: The *primary* x-rays from the x-ray tube Q reach the surface of the sample O and produce there *secondary* or characteristic x-radiation. As we shall see later on, the wave length of the secondary radiation is a function of the chemical composition of the emitting sample. The secondary x-rays emerging from the surface layer of the sample reach the concave side of the cylindrical crystal mirror. They are reflected according to their wave lengths and form a series of monochromatic x-ray images on the photographic layer F.

III. Calculation of the Positions of the X-Ray Images Along the Photographic Plate

Let us consider in Fig. 5 a cross section of the apparatus passing through the axis $A-A$ of the crystal cylinder. We shall calculate the distance $2X$ of a monochromatic x-ray image from the sample for a given wave length λ of the radiation. The direction X is coincident with the axis $A-A$. The distance $2X$ is calculated for corresponding points of sample and image. We get from the Fig. 5;

$$2X = 2 \cdot R \cdot \cot\phi$$

where R is the radius of the crystal cylinder. The glancing angle ϕ is given by the Bragg formula:

$$n\lambda = 2d \cdot \sin\phi.$$

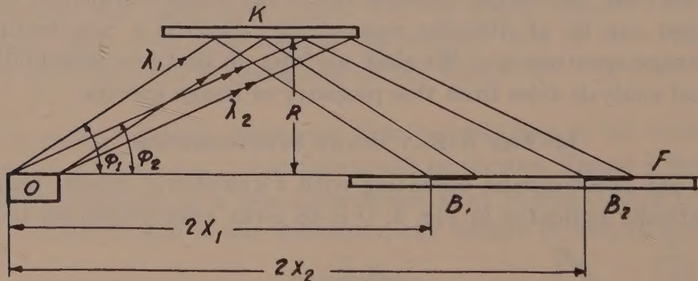


Fig. 5. Diagram of rays in the spectrograph.

By combining these formulae we get the fundamental equation of the image spectrograph:

$$2X = 2R\sqrt{\left(\frac{(2d)}{n\lambda}\right)^2 - 1.}$$

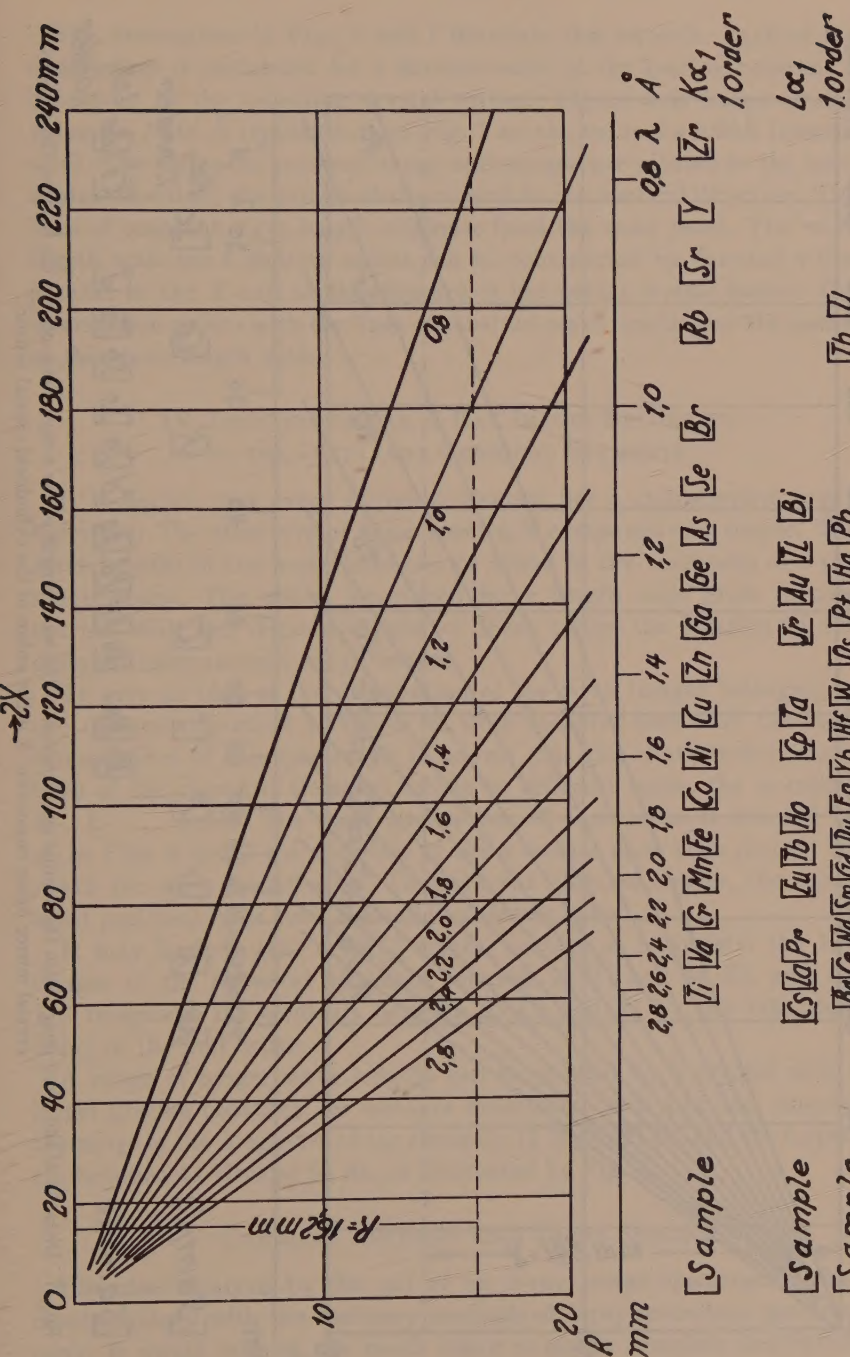


FIG. 6. Distance of the x-ray image from the sample as a function of the wave length of the x-rays given off by the sample, the cylindrical crystal mirror being NaCl. R is the radius of the cylindrical crystal mirror.

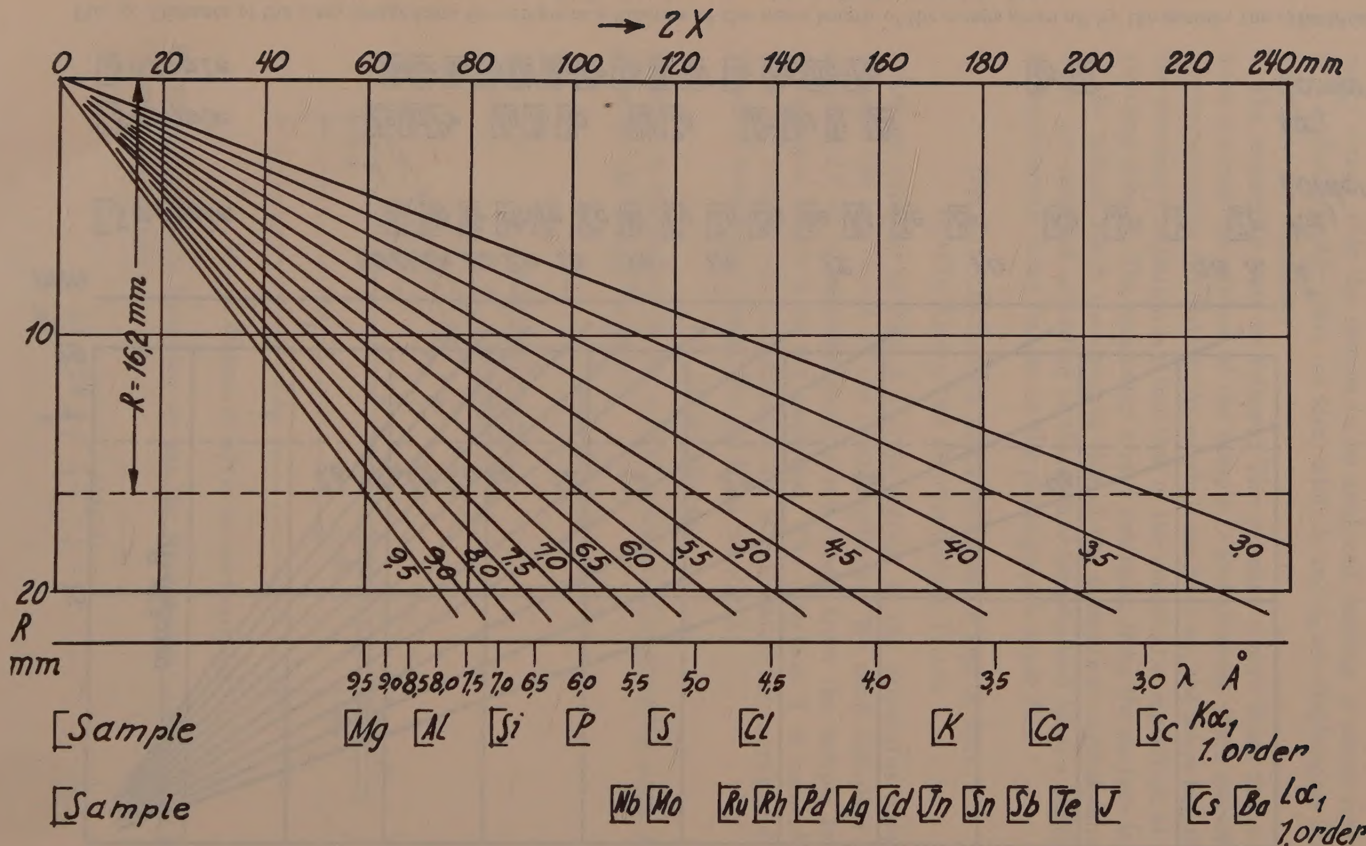


FIG. 7. Distance of the x-ray image from the sample as a function of the wave length of the x-rays given off by the sample, the cylindrical crystal mirror being muscovite. R is the radius of the cylindrical crystal mirror.

The nomograms in Figs. 6 and 7 illustrate this formula. Each of the nomograms is calculated for a definite value of the grating constant d belonging to the reflecting crystal surface. Figure 6 is based on the value $d = 2.814$ Å (rocksalt), and Fig. 7 on the value $d = 9.94$Å (muscovite). The distances between image and sample are plotted in the horizontal direction; the radius of the crystal in the vertical direction. The lines of constant wave length originate from the same point. The wave length scale for a definite radius can be constructed by drawing a line parallel to the X-axis at the distance of the actual crystal radius. The intersection points with the lines of constant wave length give the points of the wave length scale.

IV. Location of the X-Ray Images Belonging to the Different Chemical Elements

It is known that every chemical element has a characteristic x-ray spectrum. The structure of these spectra is comparatively simple. The wave lengths of the x-ray lines can be found in the textbooks of x-ray spectroscopy. The above mentioned wave length scale gives in connection with the respective data of these tables the location of the different characteristic x-ray images.

To give an idea of the distribution of the x-ray images belonging to the different chemical elements we shall consider here only the most intense lines of the spectra. It is known that the characteristic x-ray lines of the chemical elements belong to different series, the so-called, K-, L-, . . . series. The most intense line of each series is denoted by α. In Figs. 6 and 7 the positions of the α images have been plotted beneath the wave length scale. To make the diagram clearer, the calculated positions have been located in different rows.

It may be seen that a NaCl crystal enables us to register the K-α images of the chemical elements between 22 Ti and 40 Zr, and the L-α images of the elements between 55 Cs and 92 U., the reflections being of the first order.

A range of longer wave lengths can be studied by a crystal with a larger grating constant, for instance muscovite. Such a crystal supplies the range of the K-spectra of the elements 12 Mg to 21 Sc, and the L-spectra between 41 Nb and 56 Ba, as illustrated by Fig. 7.

V. Chemical Analysis with Image Spectra

Chemical analysis by the aid of an x-ray image spectrograph has much analogy with the ordinary methods of x-ray secondary spectroscopy. It would require too much space to give a complete description

here of the methods of chemical analysis. We can consider only some characteristic features of the new method.

It is known that the characteristic x-radiation of a sample depends only on its elementary components. The molecular structure does not alter the wave length of the x-radiation emitted. The intensity of the secondary radiation is determined by the concentration of the chemical element in question. In ordinary x-ray spectroscopy the sample must be made homogeneous by chemical or mechanical processes. The analysis gives only the average content of each chemical element in the sample. A characteristic property of our new method is that the sample does not need to undergo any chemical or mechanical treatment. An essential condition is however, that its surface should be sufficiently plane. The spectroscopic process does not alter the sample, so it can be conserved and the result of the analysis can be controlled whenever necessary.

As we have seen above, a primary x-ray source is needed to excite the characteristic secondary radiation of the sample. The wave length of the primary x-rays must be somewhat shorter than the characteristic radiation of the sample in question. The same rules must be followed here as in ordinary secondary spectroscopy.

After a sufficient exposure several spectral images will appear on the photographic plate. These images have different structures according to the distribution of the chemical elements in the sample. In order to make an analysis with such image spectra, first the wave length of each image must be found and second the correlation between corresponding points of image and sample surface. A general rule for the solution of this problem cannot be given as there exists a great variety of different samples.

The interpretation of an x-ray image spectrum can be made easier by comparing it with the image spectrum of another sample of known composition. As the image method does not give simple spectral lines, but images of the same form as the sample in question, a sample for comparison can be placed beside the sample during the exposure. The spectrum of the sample of comparison gives then suitable wave length marks.

The method of comparing image spectra can even be applied to quantitative analysis of samples. There can be arranged a series of known samples of comparison with graduated chemical composition. These samples produce an intensity scale on the image spectrum. Quantitative estimation of the chemical content of a sample can be done by comparing the image intensity of the sample in question with the intensity scale given by the known samples.

There are two fields of chemical analysis, where the x-ray image method is particularly suitable: first, the investigation of chemical heterogeneities in a sample and second, the analysis of samples of very small size.

As we have seen in Fig. 3 each chemical heterogeneity will appear in the x-ray spectral images. There are many problems in mineralogy, metallography, and in other branches of inorganic chemistry where, besides the total elementary analysis, the distribution of elements in small parts of a sample also needs to be studied. The methods used hitherto had a more indirect character. By the x-ray image method chemical heterogeneities can be studied with great accuracy.

The x-ray image method enables us to make spectra of samples of very small dimensions. The intensity of an x-ray spectral image is to a very great extent independent of the volume of the emitting sample. The time of exposure needed for the production of an image spectrum is therefore nearly the same foı a very small sample as for a larger one. The microscopic x-ray images can be enlarged and studied in the same way as the images of macroscopic objects. It is of very great interest that this method of spectroscopy does not destroy or alter the samples of microscopic dimensions under investigation.

VI. Examples of Image Spectra

We shall now illustrate the foregoing description of the new method with a few examples of image spectra. They have been taken by an arrangement according to the scheme outlined in Fig. 4. The primary x-ray tube was of the Coolidge type. It had a target of molybdenum and was operated at 60 kV and 5 mA. The distance between the target of the tube and the irradiated sample was about 15 mm. A cylindrical crystal mirror of rocksalt with a radius of 17.2 mm. was used.

Figure 8 shows a series of image spectra of different samples. They have been plane polished and covered with a tinfoil having a rectangular aperture according to Fig. 9. The aperture had the dimensions 4 mm.$\times$ 6 mm. On one side of the aperture a piece of brassfoil was fastened, giving suitable wave length marks (Cu- and Zn-radiation) along the spectra. The scale for the chemical elements was found by the diagram of Fig. 6. The image spectra in Fig. 8 correspond to following substances:

1. The brassfoil for comparison only (Cu and Zn).
2. Bornite (Cu K-α and K-β radiation).
3. Alloy of 12% Ni, 64% Cu and 24% Zn.
4. Alloy of 50% Co and 50% Ni.
5. Cobaltite.
6. Ore containing Fe and Zn.

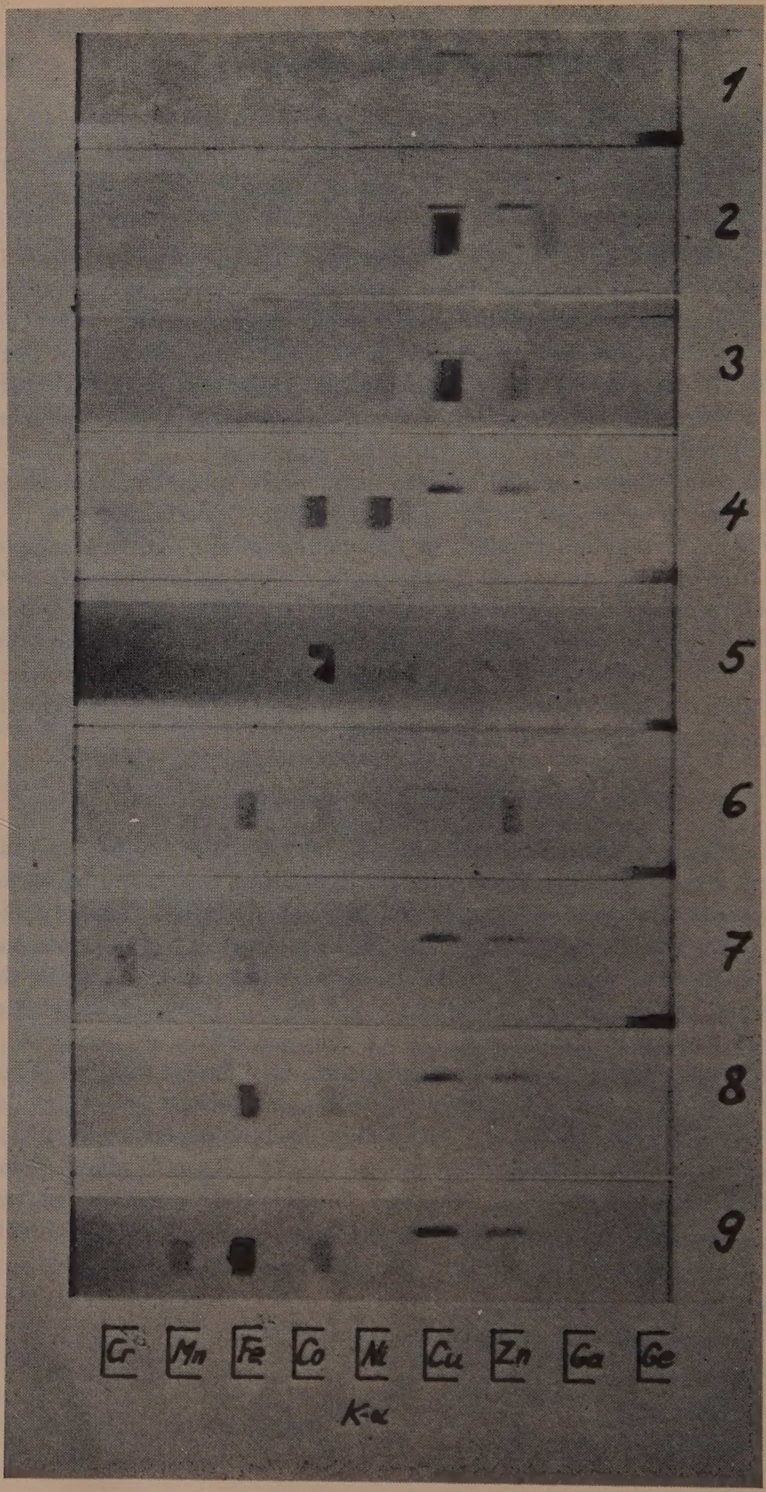


FIG. 8. Image spectra of different substances.

7. Chromite (containing Cr, Mn and Fe).
8. Ore (containing Mn and Fe).
9. Franklinite (containing Mn, Fe and Zn).

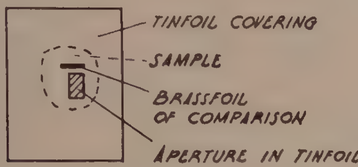


FIG. 9. Preparation of sample for spectroscopic analysis.

We see that in many cases even the K_β images appear in the spectra. In the case of homogeneous samples, as in 2, 3, 4 and 9 the spectral images have the size of the whole aperture of the tinfoil. In the other cases only the parts containing the chemical element in question will be blackened in the photograph.

Even microscopic heterogeneities in the chemical composition will appear in the spectral images. Figure 10 shows different enlarged spectral images of the same ore sample. They can be compared with the ordinary microphotograph of the same sample. The x-ray spectral images show the distribution of the elements Fe, Cu and Zn on the surface of the ore sample.

The method of x-ray image spectroscopy has been developed at the Riksmuseum of Natural History in Stockholm. The author would like to express his grateful thanks to Professor G. Aminoff, head of the Mineralogical Department, for giving him the opportunity to carry out this investigation.

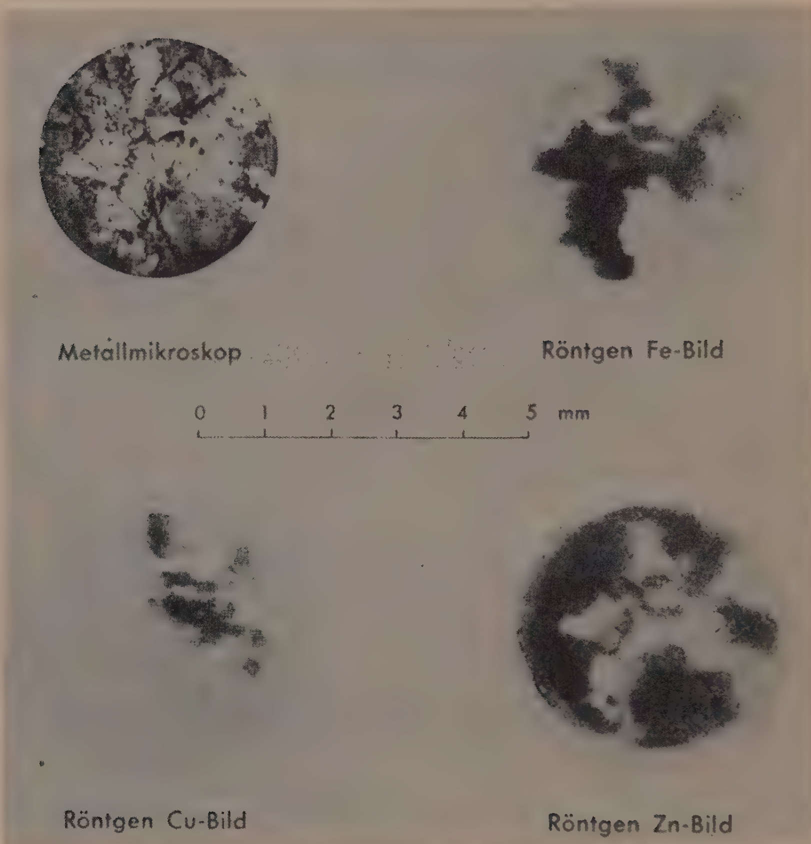


FIG. 10. Distribution of chemical elements in an ore-sample.

PETROLOGIC RESULTS OF A STUDY OF THE MINERALS FROM THE TERTIARY VOLCANIC ROCKS OF THE SAN JUAN REGION, COLORADO

E. S. Larsen, John Irving, F. A. Gonyer and E. S. Larsen, 3rd

Harvard University, Cambridge, Mass.

(*Continued from p. 905, (August), 1937*)

7. THE PLAGIOCLASE FELDSPARS

Esper S. Larsen and John Irving

Contents

Occurrence

Plagioclase is the chief mineral among the phenocrysts of the San Juan lavas. In the basalts, plagioclase phenocrysts are generally absent; in the pyroxene andesites they are almost invariably present but in

moderate amount; they are present in nearly all the rocks between andesite-latite and latite-rhyolite. They decrease in amount in the rhyolites, and some of the silica-rich rhyolites carry no plagioclase phenocrysts at all.

In most of the rocks the phenocrysts are fairly uniform in size and average about 3 to 5 millimeters in length. Exceptional rocks from nearly all the formations have much larger crystals—up to or exceeding 20 millimeters in length. The Fisher and Los Pinos latite-andesites and some of the Hinsdale andesite-basalt commonly have large phenocrysts.

In the basalts and pyroxene andesites the plagioclase phenocrysts form thin plates, but in the more siliceous rocks they occur in thicker tablets.

Method of Determining the Plagioclase

In this study the plagioclases were determined in thin section by the usual methods. In some of the zoned feldspars the Fedorov stage was used. In every specimen the final and most reliable determination was made by measuring the indices of refraction by the immersion method. For rocks with abundant plagioclase, a part of the powdered rock was used, otherwise the feldspar was concentrated by removing most of the femic minerals and the groundmass with an electromagnet or a heavy liquid. The determinations are believed to be correct to within about $\pm 3\%$ of the anorthite content.

Variation in the Composition of Plagioclase in a Single Rock

Types of Variation

Homma[32] has given an excellent and detailed classification of the types of zoning in plagioclase. Nearly all of Homma's types have been found in the San Juan lavas, but the commonest are the three varieties of wavy-oscillatory; namely, irregular-oscillatory, normal oscillatory and non-oscillatory. In most of the specimens there is considerable variation in the types of zoning, and even a single crystal may show some variation in different parts. However, unless the rock has two distinct feldspars, there is some system in the variation, and much of the lack of uniformity is due to absence of a core or a zone, or due to the thickening or thinning of zones, or the manner in which the crystal is cut by the thin section. Two types of multiple structures in which there is a change in the composition, usually accompanied by resorption, are also common in San Juan lavas.

The following is a simplification of Homma's classification of the varia-

[32] Homma F., The classification of the zonal structure of plagioclase: *Mem. College Sci., Kyota Imp. Univ.*, Ser. B., vol. **11**, pp. 135–155, 1936.

tions in the plagioclase of a single rock, which for the present purpose seems desirable.

Even (with little variation in composition)

The calcic phenocrysts of some of the basalts and pyroxene andesites are rather uniform in composition, except for a thin sodic envelope. Such phenocrysts commonly have "checkerboard" inclusions of the groundmass.

The plagioclase crystals of some of the granular rocks are nearly homogeneous as are those in many of the basalts which lack plagioclase phenocrysts. Also the plagioclase in the groundmass of some of the other rocks possess fairly uniform composition.

Normal

Feldspars that become more sodic from center to border are the main type found in the San Juan lavas, but they usually show an irregular gradation with some interruptions and oscillations.

Reverse

Feldspars that become progressively more calcic from the center to the border are rare in the San Juan lavas.

Oscillatory

Homma subdivides oscillatory zoning into 15 types. Most of his types are present in the San Juan lavas. Probably the commonest type, and the type that shows the most conspicuous oscillatory zoning, is Homma's wavy-oscillatory-even, in which the variations are similar to sine curves with little change in average composition from center to border. One crystal shows 30 rather even oscillations. The maximum difference in composition between different zones is seldom over 10 per cent anorthite. Commonly there are one or more very thin layers that are much more calcic than the others. Oscillatory zoning appears to be most common and well developed in the andesitic rocks and in feldspars with compositions ranging between An_{35} and An_{50}. It was not found in bytownite.

Calcic Cores (multiple promoted)

Feldspars of this type, with calcic cores which have sharp boundaries in contact with the more sodic outer part, are common but are not easily distinguished from feldspars with gradational zoning. The calcic cores are usually rounded and irregularly embayed as a result of resorption before the outer sodic zone was deposited. In some crystals only a

skeleton of the original calcic core is left. Of the seventy rocks examined that contain plagioclase phenocrysts, eight show conspicuously this type of zoning with differences between the core and outer shell of from 11 to 31 per cent in anorthite content. Many of the other rocks show a less conspicuous zoning of this type.

Figure 15*a* shows a feldspar of this type from the Los Pinos andesite (Con. a) of Green Ridge in the Conejos quadrangle. In the diagram to the right of the photograph, the composition of the zoned feldspar is plotted from the center to the border of the crystal according to Homma.[33] The core of this crystal is composed of a fine irregular mixture of two feldspars of very different compositions. The more sodic one occurs in patches and is replacing the more calcic portion. The diagram showing change in composition is diagrammatic for the core as the intergrowths are too fine to show on the scale used. Only a few of the plagioclase phenocrysts of this rock show calcic cores.

Sodic Cores (*multiple retarded*)

These feldspars have sodic cores and sharply separated more calcic outer zones. The less complex of these show a clear core of sodic composition with a rounded or embayed outline; a clouded intermediate zone, whose outer border is likewise rounded or embayed; and an outer zone that is clear and calcic in composition. The intermediate clouded layer in some crystals extinguishes approximately with the core, in others, approximately with the outer zone, and in some crystals it gives a moderate extinction with either the core or the border and a slight extinction with the other. This cloudy zone is a very fine intimate mixture of the groundmass, the core, and the border feldspars.

Illustrations of this type of zoning are shown in figures 15*b* and 16. Figure 15*b* represents a feldspar in a quartz basalt from a flow in the Santa Fe formation in Santa Clara Creek, near Espanola, New Mexico. Nearly all the phenocrysts of this rock show somewhat similar zoning. In this crystal the core is clear; the clouded zone contains some groundmass, and abundant patches sufficiently large to give clear extinctions that are a little more calcic than the core. Some less distinct patches are present that have nearly the composition of the outer rim. Figure 16 shows three feldspar crystals from the Antone quartz basalt of Cerro Ortez, New Mexico (NM 213). Nearly all the plagioclase phenocrysts of the rock show similar zoning. The rock has a few large grains of sanidine. In 16*a* the clouded zone is a mixture of groundmass and a feldspar of composition near that of the core, but some approaches that of the calcic

[33] *Loc. cit.*

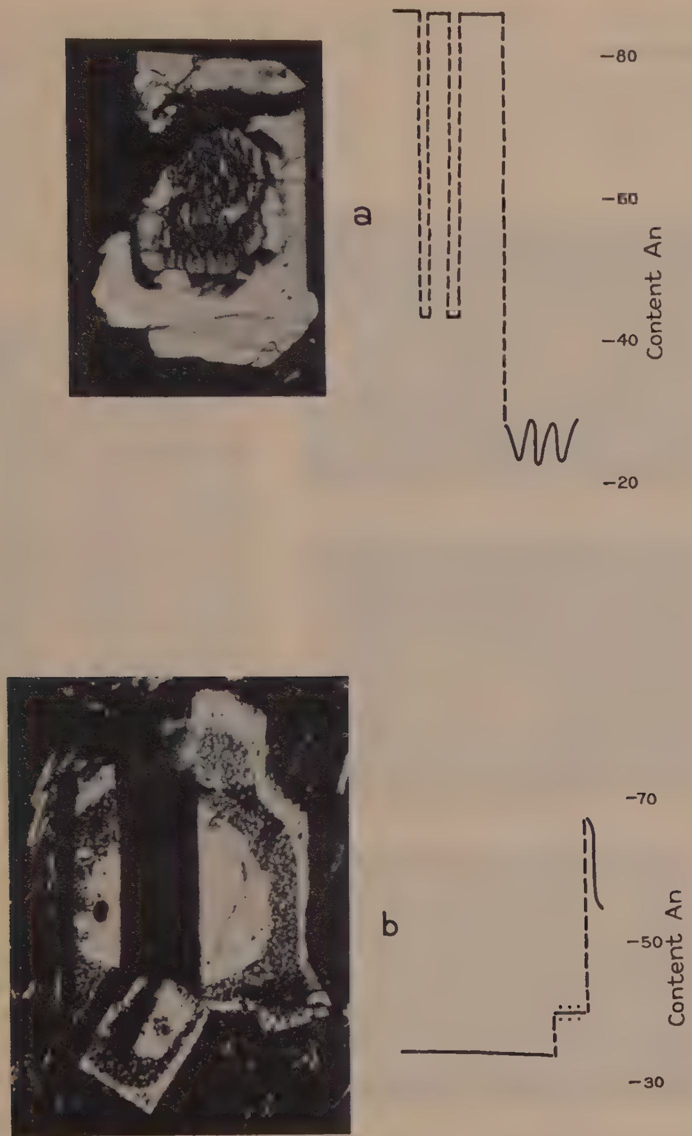


FIG. 15. Zoning in plagioclase. Photographs ×55, plots ×110.
(a) Los Pinos andesite. Shows calcic cores partly replaced by more sodic feldspar.
(b) Quartz basalt. Shows sodic cores.

part of the outer border. In 16*b* the clouded zone is made up chiefly of a feldspar more calcic than the core. In 16*c* the mottled central area is composed of a coarse mixture of groundmass, clear plagioclase like that of the calcic inner part of the clear border, and some thin layers next to the groundmass that are more sodic.

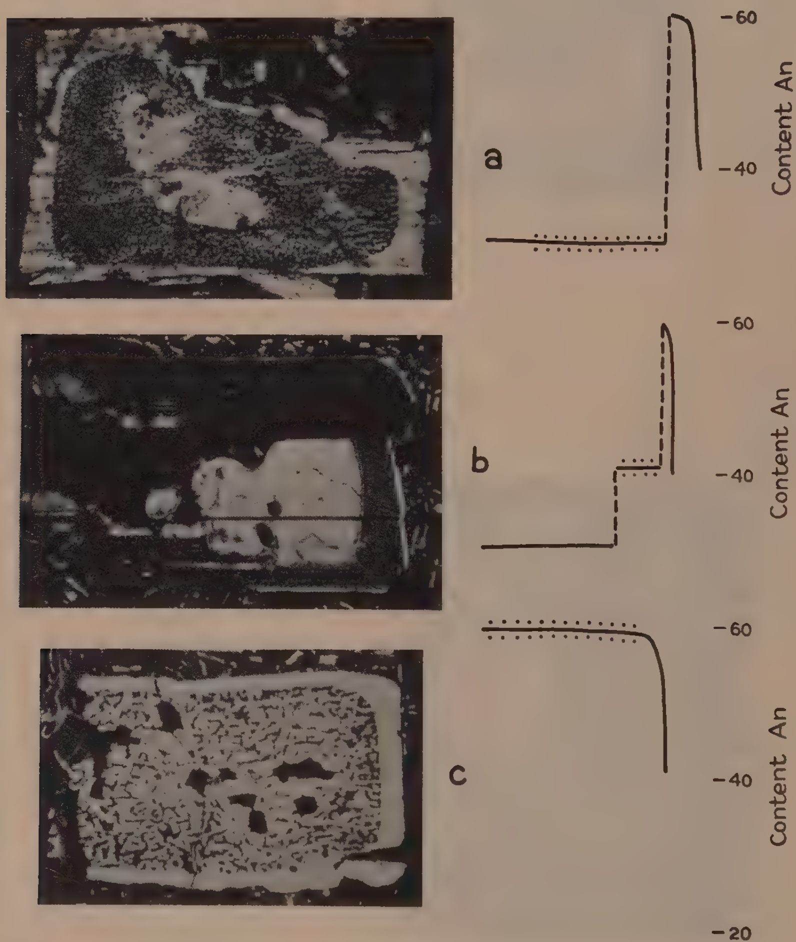


FIG. 16. Zoned plagioclase from Antone quartz basalt. Photographs and plots, ×60.

Feldspars with this type of zoning, with differences in anorthite content of the two zones ranging from 12 to 30 per cent were found in seven of the seventy rocks included in this study. They are especially common in the quartz basalts but were found likewise in some andesitic rocks.

Some of the phenocrysts show a multiple structure with two or more sharp breaks, usually with resorption. One such phenocryst from the Conejos andesite (Con 320) is shown in figure 17. It contains a small central area of An_{60}; around this is a broad band with oscillatory zoning averaging An_{46}; next is a rather narrow band with some zoning and composition of An_{33}. The outline of this latter zone is rounded by resorption. Outside this is a rather narrow zone that grades from An_{56}, near the inner portion, to An_{45} at its border. The four main zones are all rather sharply bounded.

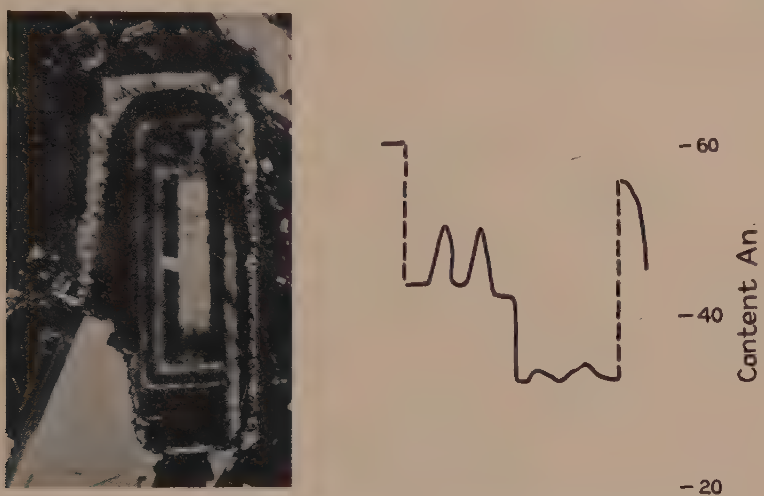


FIG. 17. Zoned plagioclase from Conejos andesite. Photograph ×70, plot ×175.

Two Types of Feldspar

Fifteen per cent of the San Juan lavas, varying from andesites to rhyolites, show two distinct types of feldspar phenocrysts. A good example is the obsidian from No Agua volcano, New Mexico. This glass has only a half per cent of feldspar phenocrysts, about a millimeter across, of which approximately 95 per cent are albite-oligoclase (An_{11}) and 5 per cent oligoclase-andesine (An_{30}). The two are in separate crystals and both show slight zoning. A Piedra quartz latite (Lag 1221) has some crystals of An_{29} and a smaller number of An_{46}. Both show moderate zoning. Conejos andesite (Con 591) has some feldspar with thin albite twin lamellae, very fine recurrent zoning without much change in composition, averaging An_{62}; also another feldspar with broad albite lamellae, large cores of An_{28} and wide borders of An_{41}. Some phenocrysts of this latter type contain remnants of a core of An_{36} enclosed in the An_{28}. A

Fisher quartz latite (Lag 2137) with phenocrysts of sanidine up to 30 millimeters long, and smaller ones of plagioclase, quartz, biotite, and hornblende, has two kinds of plagioclase phenocrysts. Some have moderate zoning (± 5 per cent anorthite) and average An_{55}; others have broad gradational zoning with cores of An_{40} and wide borders of An_{20}. Also a layer of sodic plagioclase encloses the rounded and resorbed sanidine crystals.

Feldspars near andesine in composition show more prominent zoning, especially of the oscillatory and complex types, than do those of more sodic or calcic types. Quartz latites commonly have broken crystals of oligoclase which have the normal type of zoning without much change in composition, except where calcic cores are found.

Chemical Analyses

F. A. Gonyer has made complete analyses of four plagioclases from the San Juan lavas and partial analyses of five others. These are listed in Table 5. All of the analyzed feldspars showed some zoning. None of the analyses shows an appreciable deficiency in SiO_2 compared with the theoretical composition, but those of NM 3004 and SC xx show an excess of 2.4 per cent SiO_2, and NM 3 has an excess of 4.6 per cent.

All the plagioclases contain appreciable amounts of K_2O; the amount is low in the calcic members but increases with the soda content. In terms of orthoclase, the calcic labradorites have about 1 part of orthoclase to 18 parts of albite plus orthoclase, whereas the andesines have from 1 part to 5, to 1 part to 10, averaging about 1 part to 7; the oligoclase-andesines have 1 part to 9; and the albite-oligoclase has only 1 part to 14.

In the table the analyses and the composition of the feldspar calculated from them are followed by the average anorthite content of the plagioclases and the range of the zoned feldspar, as determined by the immersion method on part of the same powder used in the analyses. The check with the chemical analyses is surprisingly close; in only one case was the deviation as much as 4 per cent, although the minerals were zoned and the averages had to be estimated. Determinations in thin sections using the following methods: (1) Carlsbad and albite twins, (2) normal to (010) and (001), (3) parallel to (010), (4) maximum extinction, and others, checked nearly as closely.

TABLE 5. ANALYSES OF PLAGIOCLASES

	1	2	3	4	5	6	7	8	9
SiO_2		53.18			59.07		59.20		65.81
Al_2O_3		28.81			25.33		24.55		20.86
Fe_2O_3		1.15			0.26		1.40		0.65
FeO									n. d.
MgO		none					0.14		none
CaO	13.02	12.01	9.15	6.01	8.60	6.71	7.09	5.31	2.19
Na_2O	3.65	4.11	5.30	6.86	6.24	6.01	5.97	6.69	8.98
K_2O	0.22	0.32	1.03	1.22	0.91	2.12	1.76	1.14	1.12
H_2O+		0.29			none		0.30		0.15
		99.87			100.41		100.41		99.76
Composition in weight per cent calculated from analyses to 100% feldspar									
Or	1	2	6+	8−	5½	12½	11	8	7
Ab	32	36	46½	61	52+	53	52½	63	81½
An	67	62	47½	31+	42+	34½	36½	29	11½
Anorthite content from optical properties									
Anorthite average	69	60	47	28	40	35	40	27	10½
Anorthite range	72	61	50	41	38	39	26	17	
	64	57	40	19	42	29	52	40	

1. Plagioclase from the groundmass of the Antone Peak basalt (NM 5) from Buffalo Butte, N. Mex. The rock has no plagioclase phenocrysts.
2. Plagioclase from the groundmass of the Hinsdale basalt (NM 402), railroad cut near Servilleta Plaga, N. Mex. The rock has no plagioclase phenocrysts.
3. Plagioclase phenocrysts from Sheep Mountain andesite (SV 9) from Summitville quadrangle.
4. Plagioclase phenocrysts from Alboroto quartz latite (Up 413), Uncompahgre quadrangle.
5. Plagioclase phenocrysts from Los Pinos latite-andesite, Conejos quandrangle, north of Green Ridge and near Goat Ranch.
6. Plagioclase phenocrysts from Alboroto latite (DN 2017), Del Norte quadrangle. Quarry 4 miles southeast of Del Norte.
7. Plagioclase phenocrysts from Piedra quartz latite (SC xx glass base analyzed), eastern part of San Cristobal quadrangle, near road north of Rio Grande and north of mouth of Spring Creek.
8. Plagioclase phenocrysts from Alboroto tridymite rhyolite (Up 3027). Uncompahgre quadrangle, east of Pine Creek, along road to Pine Creek Mesa.
9. Plagioclase phenocrysts from Hinsdale obsidian of No Agua volcano (NM 3) just east of No Agua, N. Mex. The rock has about ½ per cent of plagioclase phenocrysts, about 1 millimeter across and no other phenocrysts. 95 per cent of the plagioclase is of the type analyzed, with $\alpha=1.534$, $\beta=1.539$, $\gamma=1.542$, and 5 per cent has $\alpha=1.543$, $\beta=1.547$, $\gamma=1.551$ and has the composition An_{30}. Neither type shows much zoning and no crystals of intermediate composition were found.

Statistical Study

Relation between Composition of Rock and its Plagioclase

A statistical study of the plagioclase phenocrysts, plotted against different characteristics of the rocks in which they occur, has been made, and some of these plots are reproduced in this paper. Plagioclase phenocrysts from 70 analyzed rocks are included in the plots.

In figure 18 the anorthite content of the phenocrysts is plotted against the composition of the rocks (rhyolite to basalt) in which they occur; and in figure 19 the composition of the anorthite content of the phenocrysts is plotted against that of the groundmasses.[34]

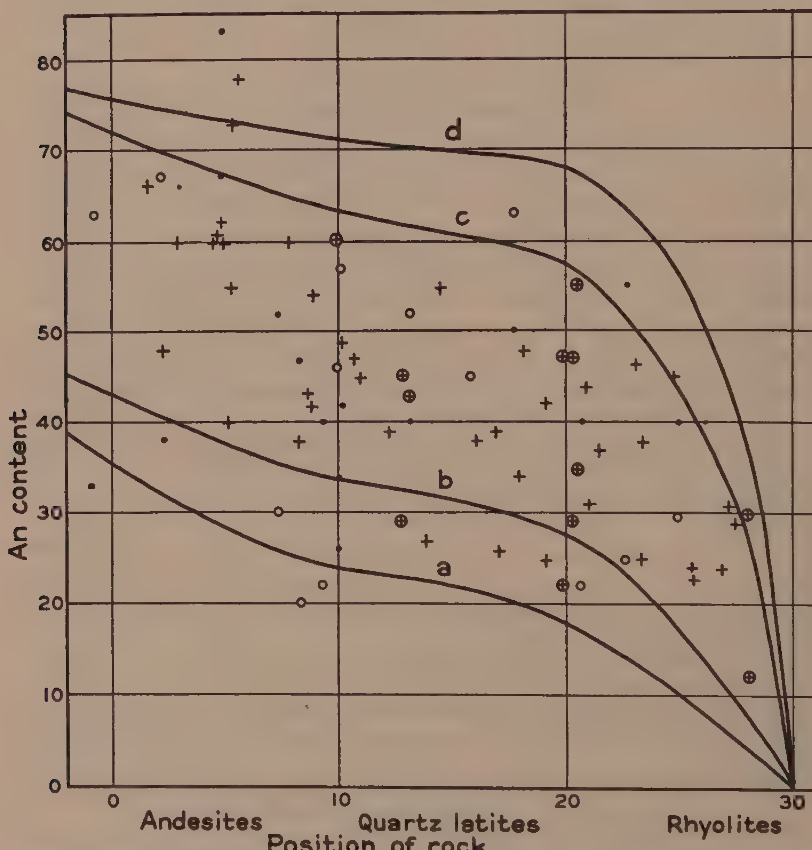


Fig. 18. The anorthite content of the plagioclase phenocrysts plotted against the compositions of the rocks in which they occur.

[34] The groundmasses were calculated from rock analyses and the phenocrysts by Rosiwal determinations.

The crosses represent phenocrysts with ordinary zoning; the dots represent the cores, and the circles the outer shells of feldspars with sharp breaks, and resorption between the cores and the outer rims; the circled crosses represent one of two different feldspars in a rock. Where one of these has cores it is represented by a dot and a circle.

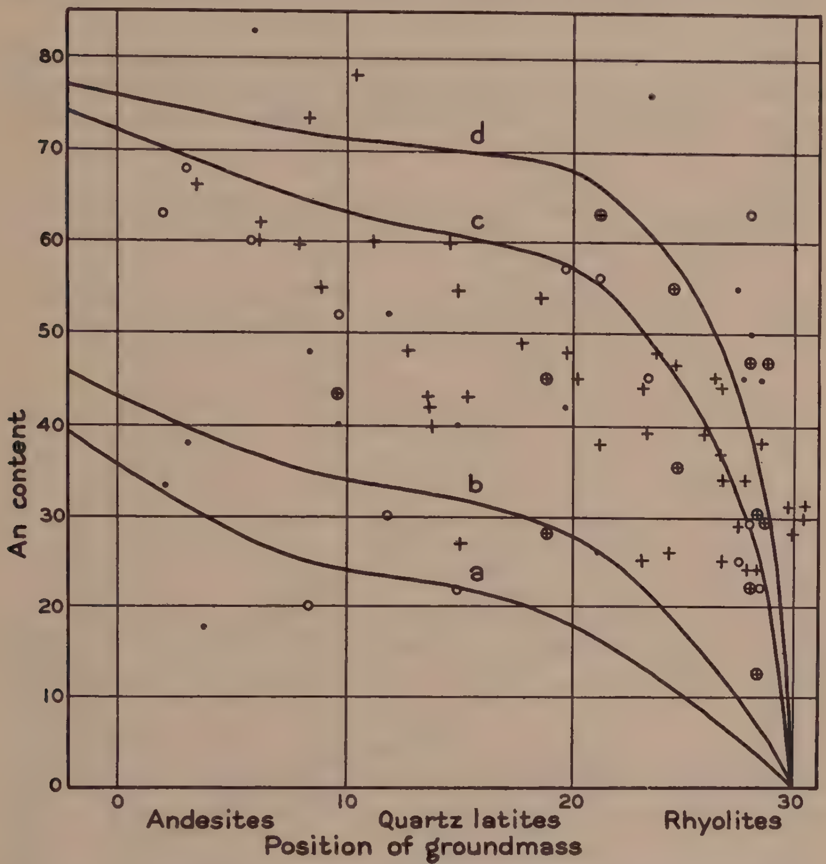


FIG. 19. The anorthite content of the plagioclase phenocrysts plotted against the compositions of the groundmasses. Conventions as in figure 18.

In these figures, the lower curve (a) represents the anorthite content of the normative feldspar $\left(\frac{\text{an}}{\text{or}+\text{ab}+\text{an}}\right)$ of the rock or groundmass; the second curve (b) represents the normative anorthite of the plagioclase $\left(\frac{\text{an}}{\text{ab}+\text{an}}\right)$; the third curve ($c$) represents the plagioclase in equilibrium

with a liquid which has the anorthite content of this feldspar (curve *d*), as shown by Bowen's[35] equilibrium diagram for the plagioclases (adding the or to the ab); the upper curve (*d*) represents the feldspar in equilibrium with a liquid with the anorthite content of the plagioclase (curve *b*).

Figure 20 is plotted in the same way as figure 18, but only those plagioclases were plotted that were believed from the thin-section study to have crystallized from the liquid which solidified to form the rock of which they are a part.

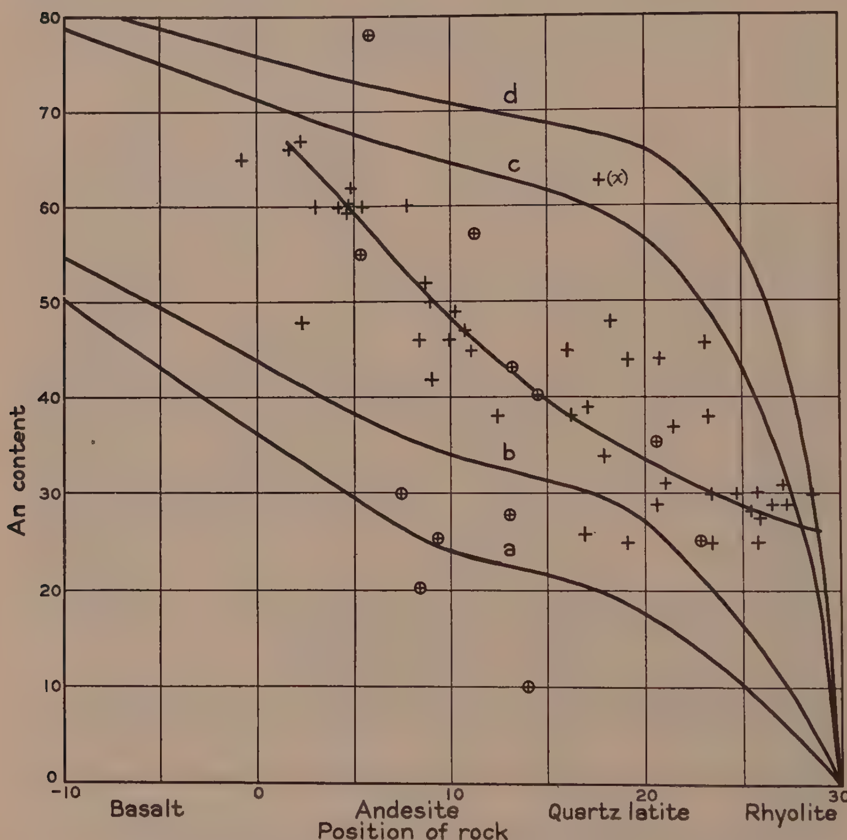


FIG. 20. Plagioclase phenocrysts that appear to have crystallized from the melt that solidified to form the rock in which they are now found, plotted against the composition of that rock. The circled crosses are less reliable than the crosses.

[35] Bowen, N. L., The melting phenomena of the plagioclase feldspars; *Am. Jour. Sc.*, vol. **35**, p. 583, 1913.

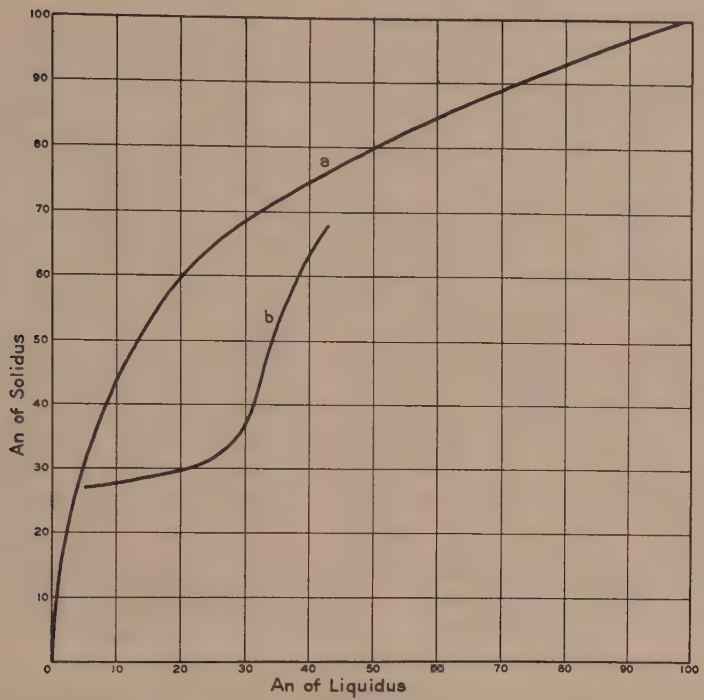


FIG. 21. Plagioclase in equilibrium with melts of various compositions. Curve *a* is constructed from the melting curves of Bowen, curve *b*, from the data of the San Juan lavas (figure 20).

Figure 21 is a plot showing the plagioclase in equilibrium with melts of all proportions of Ab and An, according to the equilibrium diagram (curve *a*), and according to the curve of crystallization for the San Juan lavas as shown in figure 20 (curve *b*). In figures 22 and 23 the anorthite contents of the plagioclase phenocrysts are plotted against the anorthite contents of the normative plagioclases and feldspars, respectively, of the rocks in which they occur. Figure 24 shows the relation between the anorthite content of plagioclase phenocrysts and the anorthite content of the normative plagioclase of the groundmasses, and figure 25 the relation of anorthite content of phenocrysts to the anorthite in the total normative feldspars of the groundmasses. In the last four figures, the diagonal line represents the anorthite content of the feldspars of the rock or the groundmasses, and the curve represents the anorthite content of the feldspar in equilibrium with a liquid of the composition of the normative feldspar.

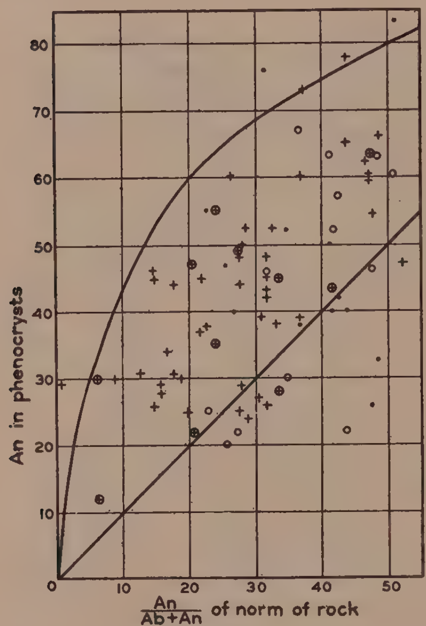


FIG. 22

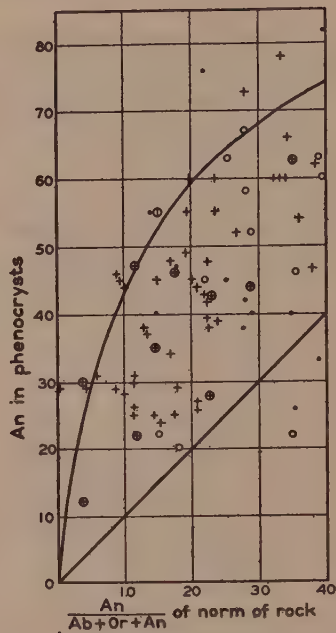


FIG. 23

FIG. 22. Anorthite content of plagioclase phenocrysts plotted against anorthite content of normative plagioclase (an/an+ab) of the rocks in which they occur. Conventions as in figure 18.

FIG. 23. Anorthite content of plagioclase phenocrysts plotted against anorthite content of the normative feldspar (an/an+ab+or) of the rocks in which they occur. Conventions as in figure 18.

The plots show that the plagioclase phenocrysts range from An_{11} to An_{83}, but by far the greater part are between An_{22} and An_{67}. About half of the rocks, whose normative feldspar has less than 13 per cent An, carry plagioclase phenocrysts, but nearly all the rocks with more calcic normative feldspar have them, excepting the basaltic types.

No satisfactory curve can be drawn to show the relation between the composition of the plagioclase phenocrysts and the composition of the rock, or the groundmass, or the anorthite content of the feldspar of the rock or groundmass. For all such curves the rocks or groundmasses of about the same compositions have phenocrysts that vary from 30 to 50 per cent or more in anorthite content.

In figure 20 that portion of the plagioclase phenocrysts that appear to have crystallized from the magma of the rock in which they were finally frozen is plotted against the composition of the rock, and in this plot there is sufficient clustering of points to justify drawing a curve. The point that deviates most (marked x) represents a rock that has 1.4 per

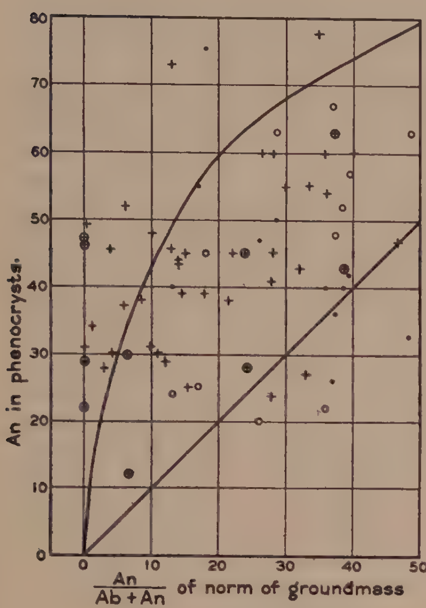


FIG. 24

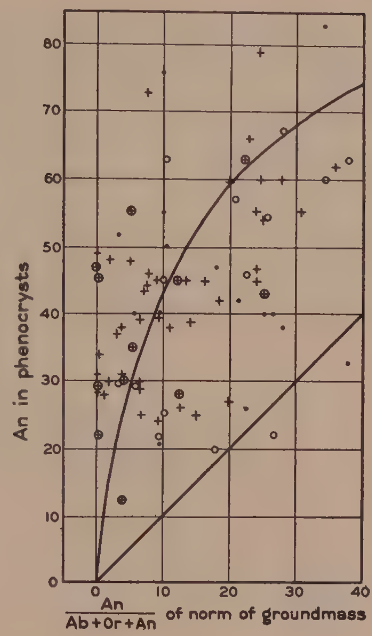


FIG. 25

FIG. 24. Anorthite content of plagioclase phenocrysts plotted against the anorthite content of the normative plagioclase (an/an+or) of the groundmasses of the rocks in which they occur. Conventions as in figure 18.

FIG. 25. Anorthite content of plagioclase phenocrysts plotted against the anorthite content of the normative feldspar (an/an+ab+or) of the groundmasses of the rocks in which they occur. Conventions as in figure 18.

cent less Na_2O than other rocks of about the same composition and is therefore abnormal. The feldspars used in figure 20 were plotted against the composition of the plagioclase and of the total feldspar of the rock in which they are found. They fall in a broad zone but yield a poorer curve than that shown in figure 20.

Points are clustered at the extreme rhyolite end, at about 24 to 30 per cent of anorthite, and this appears to be the lowest anorthite content of phenocrysts that is present in appreciable amounts. The extreme outer zone of many of the plagioclases is more sodic but this portion is present only in very small amount, as the total percentage of plagioclase phenocrysts in these rocks is less than ten.

In most of the plots nearly all the phenocrysts fall between the curves of the anorthite content of the normative feldspar (or plagioclase) of the rock or groundmass, and of the feldspar in equilibrium with a liquid of that composition. The curve for the plagioclase phenocrysts in equilib-

rium with the groundmass feldspar, obtained by plotting phenocrysts against anorthite content or total feldspar of the groundmass (Fig. 19), comes nearer to fitting the plotted points than any other.

The relation between the composition of the rocks and groundmasses, and the proportion and composition of the plagioclase phenocrysts is shown in Table 6.

TABLE 6. COMPOSITION AND AMOUNT OF PLAGIOCLASE PHENOCRYSTS IN ROCKS AND GROUNDMASSES OF DIFFERENT COMPOSITION

Rocks		Number of rocks		Per cent of plagioclase phenocrysts		Composition of plagioclase phenocrysts	
		lacking plagioclase phenocrysts	carrying plagioclase phenocrysts	Range	Average	Range An.	Average An.
Basalts	−10 to −5	4	1	0 to 2	1	—	70
	−5 to 0	0	2	2 to 3	2	42 to 63	54
Andesites	0 to 5	2	9	0 to 40	11	35 to 82	60
Quartz	5 to 10	0	13	1 to 38	17	20 to 78	49
Latites	10 to 15	0	11	1 to 33	16	22 to 64	43
	15 to 20	0	10	17 to 30	26	25 to 63	39
Rhyolites	20 to 25	0	14	3 to 27	14	22 to 55	38
	25 to 30	3	7	0 to 16	5	12 to 31	27
Ground-	−10 to −5	1	1	0 to 1	1	—	70
masses	− 5 to 0	3	1	0 to 2	1	—	65
	0 to 5	2	3	0 to 22	5	30 to 66	63
	5 to 10	0	7	2 to 23	6	20 to 82	63
	10 to 15	0	10	1 to 38	15	30 to 78	50
	15 to 20	0	9	8 to 28	20	22 to 60	45
	20 to 25	0	13	4 to 39	19	25 to 63	43
	25 to 30	3	21	0 to 31	14	11 to 62	35
	30 to 35	0	2	4 to 8	6	30 to 31	31

The proportion of plagioclase phenocrysts is low in the basalts, high in the andesites and quartz latites, and low in the siliceous rhyolites. Compared with the composition of the groundmasses, the plagioclase phenocrysts are low in amount where the groundmass is basaltic, rather low where andesitic, high where it has the composition of quartz latite and less siliceous rhyolite, and low where it is highly siliceous.

The average composition of the plagioclase phenocrysts changes progressively from An_{70} in the basalts to An_{27} in the siliceous rhyolites. Many of the siliceous rhyolites and basalts lack plagioclase phenocrysts.

As a further study of the plagioclase in the rhyolitic rocks, forty-four such rocks from six of the formations were studied. Most of these rocks were not analyzed, but their compositions were determined with the microscope. The rocks varied from rhyolites to quartz latites near the rhyolites. No quartz latites near the andesites were used. Five of the rocks contained plagioclase phenocrysts that were strongly zoned or were of two kinds, and these rocks were disregarded. Table 7 gives the data on the feldspar of these rocks.

TABLE 7. DATA ON PLAGIOCLASE PHENOCRYSTS IN RHYOLITIC ROCKS

Per cent of sanidine in feldspar phenocrysts	Number and kind of rocks	Anorthite content of plagioclase phenocrysts	
		Range	Average
100	3 rhyolites	—	—
99 to 66	7 rhyolites	29 to 33	30
	2 rhyolites	—	7[a]
66 to 33	4 rhyolites	24 to 35	27
	1 rhyolite-latite		34
33 to 1	3 quartz latites	25 to 35	30
	5 rhyolite-latites	25 to 47	36
0	1 quartz latite		40
	5 rhyolite-latites	29 to 50	40
	7 rhyolites	24 to 45	33

[a] One of these is a Willow Creek rhyolite from near the mines at Creede. It is somewhat altered and may have been albitized.

This table confirms the conclusion, drawn from the smaller amount of data obtained from the analyzed rocks, that in the extreme rhyolites, and especially in the rocks with much more abundant phenocrysts of sanidine than of plagioclase, the plagioclase has the composition An_{30}. Most of these rocks have less than 10 per cent of phenocrysts and less than 2 per cent of plagioclase. They have less than 1.2 per cent of CaO and only a few per cent of normative anorthite. To this group belong the rocks called "cavernous rhyolite" in the field. They are present in, and form an appreciable part of, all three of the rhyolitic members of

the Potosi series and form widespread, thick flows. In many places they are at the base of the rhyolitic formations and filled in the valleys and other irregularities of the surfaces on which the formations spread.

Plagioclase in Relation to Other Phenocrysts

Table 8 shows the result of taking into consideration the phenocrysts associated with the plagioclase.

TABLE 8. RELATION BETWEEN PHENOCRYSTS IN LAVAS AND CHARACTER OF PLAGIOCLASE

Phenocrysts	No. of rocks	No. with multiple zoning		No. with two plagioclases	Average composition of plagioclase phenocrysts	Remarks
		(calcic cores)	(sodic cores)			
Olivine (no quartz)	4				Calcic	Plagioclase phenocrysts absent in most.
Olivine+quartz	5	2	2		Outer zone mostly calcic	
Pyroxene ±biotite	2	1			Average	
Pyroxene +hornblende ±biotite	17	1	3	2	Calcic	
Pyroxene +biotite	9	1			Average	
Hornblende +biotite	6	1		3	Sodic	
Biotite	13	3		2	Average to sodic	
Quartz and orthoclase	12		1	3	Average to sodic	Plagioclase phenocrysts absent in some.
Orthoclase ± quartz	17	1	1	5	Average	Plagioclase phenocrysts absent in some.

Amount and Composition of Plagioclase

Table 9 shows the relation between the amount and composition of the plagioclase phenocrysts in the lavas. The plot shows that calcic andesine is the most abundant and that plagioclase phenocrysts less calcic than An_{20} are very rare, while those more calcic than An_{70} are in small number.

The rocks of the Antone andesite and Hinsdale andesite-basalt, which are basaltic in habit, are very low in plagioclase phenocrysts, but the rocks of the other formations have varying amounts and none are especially high or low.

TABLE 9. RELATION BETWEEN AMOUNT AND COMPOSITION OF PLAGIOCLASE PHENOCRYSTS

Composition	Number of rocks	Per cent of plagioclase phenocrysts	
		Range	Average
An_{10} to An_{20}	1	—	½
An_{20} to An_{30}	17	tr to 27	10
An_{30} to An_{40}	13	1 to 30	13
An_{40} to An_{50}	25	1 to 33	13
An_{50} to An_{60}	10	1 to 40	15
An_{60} to An_{70}	9	1 to 22	7
An_{70} to An_{80}	3	10 to 23	17
An_{80} to An_{90}	1	—	4

CRYSTALLIZATION OF PLAGIOCLASE IN ROCKS

Our most detailed knowledge of the crystallization of the plagioclase in rocks is based on Bowen's study of their melting phenomena.[36] The other constituents present in the rocks will modify Bowen's curves by an unknown amount. We know that they lower the temperature of crystallization by several hundred degrees. The potash feldspar in rocks may have an especially important influence as it enters into the plagioclase crystals in moderate amount. Likewise, the sanidine crystals take into solution a large amount (up to 50 per cent) of soda feldspar and a small amount of lime feldspar. In the basalts and andesites the amount of potash feldspar in the rocks is small, and much of it enters into solid solution in the plagioclase. This may have much the same effect on the crystallization of the feldspar as increasing the soda content in the rock. At the rhyolite end nearly all the soda enters the sanidine in solid solution, and this appears to have a large influence on the crystallization of the feldspar.

[36] Bowen, N. L., The melting phenomena of the plagioclase feldspars; *Am. Jour. Sci.*, 4th series, vol. 35, pp. 577–599, 1913.

Figure 21 (curve *b*) shows approximately the anorthite content of plagioclases that crystallize from rocks in the ordinary lime-alkali series of the San Juan area. These feldspars are much lower in anorthite than the feldspars that would crystallize from melts with the same albite-anorthite ratio according to the plagioclase equilibrium diagram, even if we add all of the potash to the soda. The presence of mineralizers and other constituents, therefore, seems to have a large influence in reducing the anorthite content of the phenocrysts.

Plagioclase phenocrysts with less than 20 per cent of anorthite are very rare in the San Juan rocks (only 1 in 70 rocks) although those with 20 to 25 per cent anorthite are fairly abundant. Phenocrysts with 20 per cent of anorthite would, according to the melting curves of the plagioclase, crystallize from a melt with only about 2 per cent of anorthite. In the San Juan lavas plagioclase appears to have crystallized until the composition of the crystallizing plagioclase was as sodic as An_{25} and the melt was very low in anorthite. Beyond that stage plagioclase ceased to crystallize or crystallized in very small amount, and the sanidine, which began its crystallization somewhat before this, took into solid solution practically all of the material of feldspar composition that remained in the melt. Indeed, there is a strong suggestion that there is a reaction relation between oligoclase and sanidine.

Another explanation for the apparent absence of very sodic plagioclase phenocrysts is that the amount of plagioclase crystallizing in this range is very small.

Explanation of the Variations in the Plagioclase

Uniform Crystals

The uniform phenocrysts are nearly all near bytownite and are in basaltic rocks. Such crystals rarely show oscillatory or irregular zoning. A possible explanation for this is that diffusion in both the liquid and solid phases is more rapid in the calcic rocks, and hence the phenocrysts tend nearly to reach equilibrium in such rocks. Another possibility is that these calcic phenocrysts crystallized slowly from a calcic melt and settled into the magma which finally crystallized into the present rock. This is suggested by the fact that these uniform calcic phenocrysts are unusually calcic for the rocks in which they are found and they commonly have a thin layer that is much more sodic. Moreover, from the nature of the diagram for feldspar equilibrium, crystals precipitating from a calcic liquid change more slowly on crystallization of the liquid than do those crystallizing from a sodic liquid, as shown in figure 26. Such crystals should have large cores with little change in composition, fol-

lowed by intermediate zones that change rapidly in composition, and finally an outer shell of very sodic feldspar. In phenocrysts, crystallization has not proceeded to completion and a greater or lesser portion of the outer zones, shown on figure 26, would be missing.

The uniform composition of the plagioclase of some granular rocks is probably explained by slow crystallization with complete reaction and equilibrium, whereas the uniform composition of the groundmass plagioclase in the porphyritic rocks is explained by rapid crystallization without equilibrium.

Normal Zoning

The simplest case of normal zoning with calcic cores grading into more sodic borders is easily explained as the result of progressive crystallization without reaction, and both the liquid and the material crystallizing become more sodic as crystallization proceeded. Starting with the plagioclase equilibrium diagram, curves have been constructed to indicate the course of crystallization under such conditions that the precipitation of zoned crystals takes place so that at all times the liquid is in equilibrium with the outer shell of the crystals, but without reaction between the liquid and crystals. Such curves for the following initial liquids and crystals are shown in figure 26. Curves 1 to 5 were constructed from the curves of the plagioclase equilibrium diagram, assuming that the feldspar was precipitated in equilibrium with the melt and that there was no reaction. The anorthite content of the outer zone of the crystals is plotted against the weight per cent of the part that is crystalline.

	Anorthite content of initial liquid	Anorthite content of first feldspar to precipitate
Curve 1	85	95
2	72	90
3	46	80
4	20	60
5	7	40

For the conditions postulated for curve 1, the first crystals to form have the composition of An_{95}. On cooling, precipitation is rapid and the change in composition of the feldspar is slow. When the outer rim has a composition of An_{80}, 85 per cent of the original melt is crystalline. From here on, crystallization becomes gradually slower and the change in composition of the precipitating feldspar becomes very rapid, so that when it has the composition An_5, 96 per cent of the original melt is crystalline. On further cooling, precipitation of highly sodic feldspar

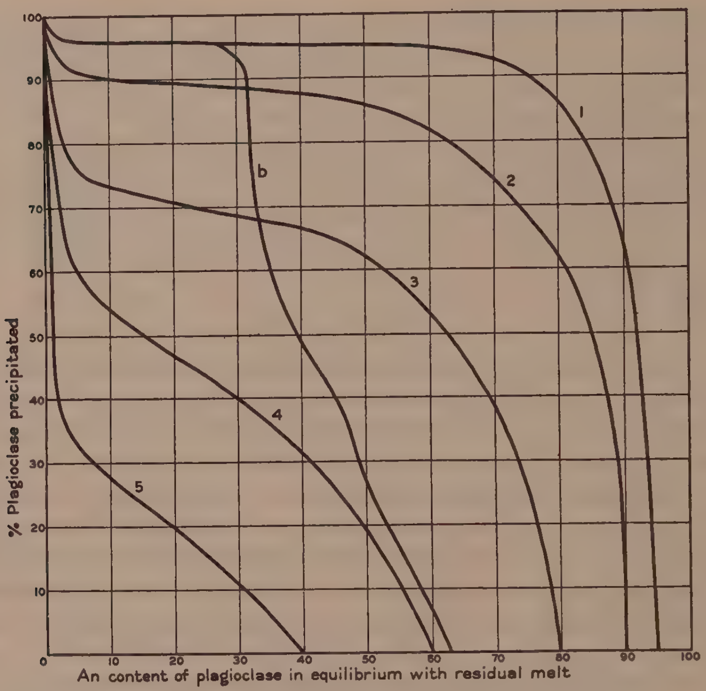


FIG. 26. Precipitation curves for progressively zoned plagioclases.

proceeds more rapidly and with less change in composition. Curves 2, 3, 4 and 5 are simply parts of curve 1 with enlarged vertical scale. Where crystallization takes place according to curve 5, the first crystals to precipitate have the composition of An_{40}. On cooling, precipitation takes place with moderate rapidity and moderate change in composition, until the precipitating crystals have about the composition An_5, when 34 per cent of the original liquid is crystalline. Here a fairly sharp break appears in the curve, and on further cooling precipitation is very rapid and the change in composition slow.

If a zoned plagioclase was formed according to curve 1, it should have a large core with only moderate change in anorthite content until the composition An_{80} is reached. At about that composition a rapid change should begin and the band including the range An_{80} to An_5 should be narrow. At about An_5 another rather abrupt change should take place and an outer zone more sodic than An_5 should be rather broad. In case the initial melt was lower in anorthite than for curve 1, the inner core would be smaller and the outer zones correspondingly larger, and if the first feldspar was more sodic than An_{80}, the rather uniform inner core

would be lacking. If a more sodic melt was used, the sodic border would become increasingly larger in amount.

Curve *b* of figure 26 represents the crystallization of a melt of composition An_{40} according to the curve of crystallization of the San Juan lavas shown in figure 20. In this curve the change in composition is more regular and less marked than in those drawn from the melting curves of plagioclase.

Figure 27 shows more clearly the kind of zoning that would result from such types of crystallization. In this figure the compositions of the different plagioclases are plotted against the distance from the center of the crystal. These curves may explain the fact that the very calcic plagioclase phenocrysts commonly show little zoning except for a narrow border of very sodic feldspar.

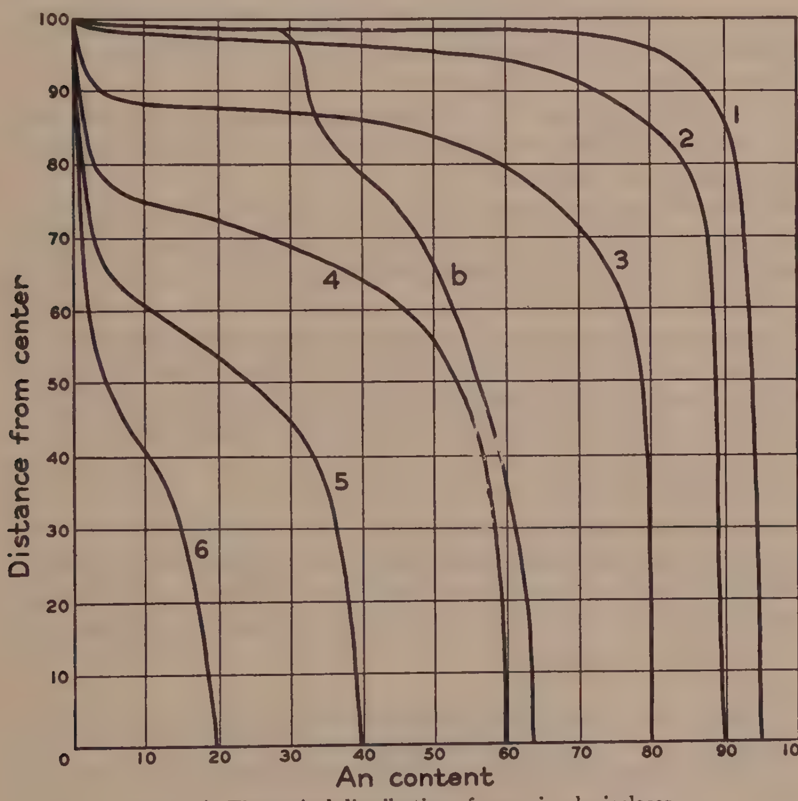


FIG. 27. Theoretical distribution of zones in plagioclases.

Curve *b* is drawn for the crystallization of a magma with 40 per cent of anorthite in the plagioclase, from the data for the San Juan lavas as shown on curve *b* of figure 26.

Among the plagioclase phenocrysts of the San Juan lavas few show zoning even approximately resembling that shown in figure 27. In addition to the progressive zoning, many of the phenocrysts show more or less recurrent zoning, others show sharp boundaries between zones, and still others have very calcic cores. Few of these phenocrysts show as wide a variation in composition as is indicated in figure 27 and few show the sodic borders. This lack of sodic borders is to be expected in liquids that did not complete their crystallization, but in a number of the San Juan lavas the anorthite molecule was almost completely removed from the groundmass as shown in figure 25. The zoning is much more like that shown on curve *b* of figure 27. These facts tend to confirm this curve.

Oscillatory Zoning

Homma[37] has classified a great many types of oscillatory zoning. In a single rock the type of oscillation varies more or less but usually within moderate limits.

For a general discussion of the subject of oscillatory zoning in the plagioclases and for references to the literature the reader is referred to the excellent paper by Hills.[38]

The explanation of regular oscillatory zones with as many as 80 repetitions requires some sort of rhythmic precipitation and Hills' theory seems very satisfactory. He thinks that crystallization starts under conditions of equilibrium. As it proceeds, by slow diffusion, the liquid next the crystal and the outer shell of the crystal itself become enriched in soda and correspondingly the liquid beyond is enriched in lime and supersaturated. Crystallization may now become slow, or may even cease, and diffusion will enrich the liquid next the crystal in lime and crystallization of an anorthite-rich zone may follow. This may be repeated again and again.

Some of the irregular oscillations are probably due to other causes, such as a change in the liquid in which the crystals are immersed by movement of the crystal, or by mixing of two magmas, or by change in conditions such as loss of mineralizers, or by movement through convection currents.[39]

[37] Homma, F., The classification of the zonal structure of plagioclase; *Mem. College Sci., Kyoto Imp. Univ.*, ser. B., vol. **11**, pp. 135–155, 1936.

[38] Hills, E. S., Reverse and oscillatory zoning in plagioclase feldspar; *Geol. Mag.*, vol. **73**, pp. 49–56, 1936.

Phemister, J., Zoning in plagioclase feldspar; *Mineral. Mag.*, vol. **23**, pp. 541–555, 1934.

[39] Homma, F., The method and the principle of delineating the composition-variation curve of a zoned plagioclase with an example: *Mem. College of Science, Kyoto Imp. Univ.*, ser. B., vol. **12**, no. 1, pp. 39–40, 1936.

Calcic and Sodic Cores

The calcic cores, where present, have in some specimens rounded outlines with long arm-like embayments; in others only a skeleton of the calcic core remains. They are sharply separated from the more sodic border (Fig. 15*a* and 17). Their general appearance suggests a solution remnant but the long projecting arms and the skeleton forms all oriented with the host are indications that they are replacement remnants.

In the crystals with sodic cores, the core is of clear sodic feldspar, surrounded by a layer that is clouded and made up of a fine-textured mixture of groundmass, feldspar near that of the core in composition, and feldspar near that of the outer zone. The outer zone is clear and calcic. The boundary between the clouded zone and the clear outer part of the crystal is sharp and rounded (see Figs. 15*b* and 16). The boundary between the clouded zone and the clear core is rounded and sharp but some prongs of the clouded zone project into the clear portion. It seems evident that a feldspar that was much too sodic to be in equilibrium with the magma was partly dissolved and later penetrated by the liquid and partly replaced by a fine aggregate—the cloudy zone.

The calcic cores have as much as 40 per cent more anorthite than the bordering feldspar, and the sodic cores as much as 30 per cent less. They clearly did not crystallize from the magmas which deposited the outer zones. They may have crystallized in the lower or upper parts of a layered magma and floated or settled into another part of the magma. This implies that the calcic cores have densities less than those of the siliceous magmas into which they floated, and that the sodic cores have densities greater than those of the liquids into which they settled. It is doubtful if either is true, and one excludes the other.

It is more likely that the calcic cores are due to contamination of a silicic magma by basaltic material, and the sodic cores are due to contamination of a relatively calcic magma by more sodic material. The contaminating material that it furnished to the cores of the plagioclase might have been a partly crystallized liquid or a rock. The fact that many basaltic rocks which have feldspars with sodic cores have phenocrysts of quartz, sanidine, or other unexpected minerals indicates this.

Two Feldspars in One Rock

The two feldspars differ by as much as 30 per cent in anorthite con-

Homma, F., Über das Ergebnis von Messungen an zonaren Plagioklasen aus Andesiten mit Hilfe des Universaldrehtisches: *Schweiz. Min. Petr. Mitt.*, **12**, pp. 345–351, 1932.

Rittmann, A., Geologie der Island Ischia: *Zeits. Vulkanologie*, vol. **6**, pp. 208–210, 1930.

Kozu, S., and Watanabe, S., The zonal structure of plagioclase phenocrysts in pumice ejected from Komagake in 1929: *Proc. Imp. Acad. Japan*, vol. **8**, pp. 379–382, 1932.

tent and they differ in habit and kind of zoning as well. In some rocks one of the feldspars appears to be high in soda for such rocks, in others low in soda. These two feldspars in one rock seem to require much the same explanation as do the feldspars with sodic or calcic cores. One of the feldspars must have come from a contaminating rock or magma.

Origin of the Phenocrysts

Review of Evidence that Crystals are Foreign

Many facts listed below lead to the conclusion that some of the phenocrysts in the San Juan lava did not crystallize from the magmas in which they were erupted and that they are in a sense foreign crystals. Yet with few exceptions many of the rock analyses fall on a very uniform variation diagram which is essentially that of the average of the rocks from rhyolite to basalt, as calculated by Daly. Indeed, for most of the rocks, if one of the major variable oxides, SiO_2, total iron, MgO, CaO, or K_2O is known, a fair analysis of the rock can be written.

1. The average of the plagioclase phenocrysts in the silicic rhyolites is An_{28}, that in pyroxene andesites is An_{60}, and for intermediate rocks they range between these values. However, in rocks with about the same composition the plagioclase phenocrysts vary 25 to 40 per cent, or more, in anorthite content. In fact, plots comparing the anorthite content of the plagioclase phenocrysts and the composition of the rocks or of the groundmasses, and plots of the anorthite content of the normative plagioclase of the rock or of the groundmass, and the normative feldspar (or+ab+an) of the rock or the groundmass, show a very wide scattering of points. They show a surprisingly small relation between composition of plagioclase phenocrysts and composition of the rock or groundmass (Figs. 18 to 25).

2. The character of the zoning in many of the feldspars and the presence of highly calcic cores in over 10 per cent of the rocks, and of highly sodic cores in 10 per cent show that there was some addition of foreign crystals.

3. Two different kinds of plagioclase phenocrysts are present in about 15 per cent of the rocks.

4. Rocks with two kinds of plagioclase phenocrysts or with very sodic or calcic cores are commonly, though not always, found in rocks that contain other unexpected phenocrysts—such as quartz or orthoclase in basaltic rocks, rounded sanidine with outer layers of oligoclase, etc.

5. As shown in a former section of this series of papers, the pyroxene phenocrysts[40] show no systematic variation in iron content in rocks rang-

[40] *Am. Mineral.*, vol. **21**, pp. 695–699, 1936.

ing from rhyolite to pyroxene andesite. Pyroxenes that have crystallized from siliceous rocks should be richer in iron than those crystallized from a pyroxene andesite.

6. Among the San Juan lavas it is not uncommon to find two lavas, or groups of lavas, with essentially the same composition but with different phenocrysts. This is illustrated by the quartz latites of the three rhyolitic members of the Potosi volcanic series. The Treasure Mountain rocks carry phenocrysts of andesine, biotite and augite; those of the Alboroto have quartz, orthoclase, oligoclase, biotite, hornblende, and sphene; and those of the Piedra have quartz, orthoclase, andesine, biotite, hornblende, augite and sphene. The quartz latites of each of the three formations have the same range in composition. Why are the phenocrysts so consistently different in these three formations? Combined with other evidence, the best explanation is that many of the phenocrysts are foreign to the rocks in which they are now present.

The foreign crystals might have come from a different part of the magma chamber by floating, sinking, convection currents, or in some other way. They might be due to the mixing of two magmas, either or both of which had suspended crystals, or they might be undissolved remnants from a rock, more or less completely solidified, that had been dissolved or reacted on by the magma.

Accumulated by Sinking or Floating

Probably a part of these foreign crystals floated or sank into their present position. However, not all of the erratic crystals can be accounted for in that way. If a layered magma be assumed, or one that changes progressively from top to bottom, it is necessary to have the lighter rhyolitic magma above the heavier basaltic magma. Since the feldspars with calcic cores ranging from An_{40} to An_{83} are in the rhyolitic and intermediate rocks, it would be necessary, if gravity separation were the cause of the movement of the crystals, to have crystals of labradorite, or even of bytownite and probably pyroxene, float from a basaltic liquid into a rhyolitic liquid. It seems reasonably certain that such crystals have a greater density than rhyolitic magmas.

Likewise, the feldspars with sodic cores ranging from An_{26} to An_{60} are in the basaltic and intermediate rocks. In some rocks they are associated with either sanidine or quartz or both. The sodic plagioclase, sanidine and quartz must have crystallized from rather siliceous magmas and they are no doubt too low in density to have sunk into andesitic and basaltic magmas.

Where two types of feldspar phenocrysts are present in a rock, one does not grade into the other. If one crystallized in place and the other reached

its position by movement due to gravity, the movement probably took place in a magma chamber that had two distinct layers of different compositions, for if the magma changed gradually we should not find such distinct crystals. The same difficulty in density relations would be present as in the crystals with calcic or sodic cores; and the two types of plagioclase, as well as the plagioclases with cores, are believed not to have accumulated by gravity. They differ from the calcic and sodic cores chiefly in that they are not resorbed to any great extent and are not bordered by feldspar of very different composition. This is probably due to the fact that the magma was erupted shortly after the phenocrysts were added.

In rocks with one type of plagioclase and no conspicuous cores, many of the phenocrysts are foreign to the rock, since in rocks of essentially the same composition there is a large range in the composition of the plagioclase phenocrysts. For instance, in rocks in which the normative plagioclase of the groundmass (Fig. 23) is about An_{27}, some have phenocrysts less calcic than the groundmass (An_{24}) and others as calcic as An_{60}. If the plagioclase crystallizing from this liquid was An_{60}, the more sodic plagioclase might have settled by gravity from a less calcic liquid above. A liquid of this type with much quartz and orthoclase, some water and only a moderate amount of FeO and MgO might have a lower specific gravity than the crystals, but it is doubtful if such crystals of oligoclase would settle into a liquid of the composition of pyroxene andesite.

Reaction with Wall Rock

Bowen[41] has discussed the reaction of magmas on inclusions. If a magma includes a crystalline rock more siliceous than the magma, the inclusion will be dissolved if the magma has sufficient superheat, but, if not, the inclusion will be dissolved with precipitation of crystals that are in equilibrium with the modified magma. The basalt (NM 213) from Cerro Ortez, near Tres Piedras, illustrates this point. This rock contains 7% of olivine phenocrysts, 1% of pyroxene, 2% of plagioclase, and less than 1% each of quartz and sanidine. The groundmass is made up of abundant small plates of laboradorite and pyroxene grains in a submicroscopic matrix.

The quartz and sanidine are much resorbed. The plagioclase phenocrysts have clear cores of An_{42}, surrounded by a cloudy zone made up of a mixture of groundmass, An_{42} and An_{63}, and an outer clear zone of An_{63}

[41] Bowen, N. L., The behavior of inclusions in igneous magmas: *Jour. Geology*, vol. **30**, pp. 513–567, 1922.

——— *The evolution of the igneous rocks*, Princeton University Press, pp. 175–223. **1928.**

(Fig. 16). The plates of the groundmass have cores near An_{60} and more sodic borders. The sodic feldspar of the cores and the sanidine and quartz are, no doubt, crystal remnants from the partial solution of a quartz latite or granodiorite. The plagioclase appears to have been dissolved at first to the rounded boundary outside of the clouded zone. During this stage there may have been superheat. Then the magma penetrated the outer shell of the rounded grain and dissolved part of the crystal and precipitated a feldspar in equilibrium with the liquid. This formed the mixed clouded zone. The outer zone then grew by precipitation and protected the core from further reaction. That this outer zone was in equilibrium with the melt is indicated by the fact that it has nearly the same composition as the cores of the groundmass plagioclase. The amount of foreign material that was required to furnish the foreign crystals need not have been great. In other rocks more was needed. The effect of this added material would have been to change the rock toward a rhyolite. The mixed rock would fit the equilibrium diagram since, except at the extreme siliceous end, the curves for the various oxides, excepting Al_2O_3, are nearly straight lines.

If a rock less siliceous than the magma is reacted on, the results are much as in the preceding case, but it will take a greater amount of superheat to bring about complete solution and without superheat a greater proportion of crystals would be precipitated.

Mixing of Two Magmas

The mixing of two magmas of different compositions, either or both of which carried phenocrysts, would give much the same effect as reaction on included fragments, but the heat and solution effects would be simpler. If the magma that carried the chief phenocrysts was present in small amount, the phenocrysts would be farther from equilibrium with the liquid than if a large amount of the magma had carried the phenocrysts.

The probability of mixing two magmas is increased by the intimate and erratic association of rhyolitic and basaltic or andesitic rocks in the area. There are local basaltic horizons interlayered with nearly all of the rhyolitic members and there are local rhyolitic rocks in some of the andesitic members. The Treasure Mountain quartz latite illustrates these relations. It is a regularly layered formation made up chiefly of rhyolitic flows and tuff beds. It is over 1,000 feet thick in the Summitville quadrangle and has four discontinuous thin lenses of dark pyroxene andesite interbedded with it. Both rhyolites and dark pyroxene andesitic lavas must have been erupted periodically from the same vent or from vents

very close together. Williams[42] has described alternate eruptions of rhyolitic and basaltic rocks from the Newberry craters of Oregon.

Conclusions

The conclusion seems inevitable that many if not most of the plagioclase phenocrysts in the San Juan lavas did not crystallize from a magma of the composition of the lavas in which they are found. Some of the crystals that crystallized in the extruded lava may have originated by settling or floating from a higher or lower layer of magma of different composition, but most of them are not believed to have accumulated in this way, but rather by the mixing of two partly crystallized magmas, or by the reaction on solidified rock.

Some of the other phenocrysts were accumulated in the same way; probably all of the quartz phenocrysts in the quartz basalts are foreign. In an earlier part of this series of papers[43] it was stated "that quartz crystallized from the basalt magma as well as did the olivine," but the conclusion that many of the phenocrysts of sanidine and plagioclase associated with the quartz basalts are foreign, leads to the conclusion that the quartz also is foreign.

The sanidine phenocrysts in the basaltic rocks, the rapakivi-like sanidine, associated with quartz phenocrysts in some of the Fisher andesite-latites are almost certainly foreign, and probably some of the quartz and sanidine phenocrysts in the quartz latites are likewise foreign.

The pyroxenes in the siliceous rocks are essentially the same as the pyroxene phenocrysts in the andesites, and it seems probable that they are foreign, since if they crystallized from siliceous rocks they would be richer in iron.

Another conclusion, contrary to the authors' former conceptions, is that there must have been some means to mix thoroughly large bodies of magmas before eruption. The uniformity of large bodies of magma that have foreign phenocrysts makes this seem inevitable. Some of these flows are very large and even great groups of flows are quite uniform. This statement as to the uniformity of flows is based on estimates of composition in the field and laboratory. A thorough and more quantitative study of some of the flows is desirable.

Detailed studies of the crystallization of magmas after the method of

[42] Williams, Howel, Newberry volcano of central Oregon: *Bull. Geol. Soc. Am.*, vol. **46,** pp. 300–302, 1935.

[43] *Am. Mineral.*, vol. **21,** p. 685, 1936.

Vogt are highly desirable and should yield important results to supplement the equilibrium diagrams worked out in the Geophysical Laboratory. They should help to bridge the wide gap between the equilibrium diagrams and the igneous rocks. In making such studies, great care must be taken to be certain that the rock studied has neither lost nor gained phenocrysts. One favorable place for such studies is in dikes and other small intrusive bodies where the course of crystallization can be followed from the chilled borders to the center of the bodies.

NOTES ON CATLINITE AND THE SIOUX QUARTZITE

Ernest L. Berg, *University of Minnesota, Minneapolis, Minnesota.*

Introduction

The Sioux quartzite, where it outcrops at Pipestone, Minnesota, contains a thin bed of red, metamorphosed shale of peculiar properties. The shale is soft and easily carved. It is commonly referred to as "pipestone" since it was formerly used by the Indians of that region in the manufacture of pipes.

The pipestone quarry was visited by white men as early as 1836 when George Catlin journeyed all the way from New York for the purpose of seeing the renowned quarry.[1] He noted[2] that

> the Indians procure the red stone for their pipes by digging through the soil and several slaty layers of the red stone to the depth of four or five feet. From the very numerous marks of ancient and modern digging, or excavations, it would appear that this place has been, for many centuries, resorted to for the red stone, and from the great number of graves and remains of ancient fortifications in the vicinity (as well as from their actual traditions) it would seem that the Indian tribes have long held this place in high superstitious estimation, and also that it has been the resort of different tribes, who have made their regular pilgrimages here to renew their pipes.

Catlin collected a sample of the red pipestone and sent it to Professor C. T. Jackson of Boston who made a chemical analysis and reported that it was not steatite but a new substance somewhat similar to agalmatolite, a stone used by the Chinese in carving. He named the pipestone *catlinite*[3] in honor of Mr. Catlin.

At this early date the Pipestone, Minnesota, quarry was the only known source of the red pipestone. Since that time, however, several reputedly similar beds have been reported in Algonkian quartzites near Devil's Lake,[4] Wisconsin; Rice Lake,[5] Wisconsin; Sioux Falls,[6] South Dakota; and elsewhere.

Of late years catlinite has not been recognized as a separate species, but has been variously classified as a silicified shale or as an indurated

[1] Winchell, N. H., Geology of Minnesota, vol. 1 of the Final Report: The Geology and Natural History Survey of Minnesota, p. 63, **1884.**

[2] Catlin, George, *Am. Jour. Sci.*, vol. **38**, p. 138.

[3] Jackson, C. T., Catlinite or Indian pipestone: *Am. Jour. Sci.*, First Series, vol. **35**, p. 388, 1839.

[4] Woodman, E. E., The pipestone of Devil's Lake, Wisconsin: *Wis. Acad. Sci. Trans.*, vol. **5**, pp. 251–254, 1882.

[5] Hotchkiss, W. O., Mineral Land Classification: Wisconsin Geology and Natural History Survey, pp. 37–38, **1915.**

[6] Winchell, N. H., *op. cit.*, p. 542.

clay. There is, however, no published study relating to the mineralogy and petrography of the catlinite bed. Its historical importance and the peculiar properties of the stone itself seem to warrant such a study. For this reason the writer visited the outcrop during the last summer and collected samples of the catlinite bed and the overlying quartzites for laboratory study.

Field Relations

The pipestone quarry lies about one half mile north of Pipestone, Minnesota, in the southwestern part of the state. The region is mostly overlain by glacial drift but a ridge of quartzite, a mile or more in length, rises out of the prairie at the quarry site. The ridge presents a west-facing escarpment 25 to 30 feet high (see Fig. 1). A modern rock quarry which started at the face of the escarpment has been worked back into the ridge. The quarry walls offer a fresh vertical section for study. The Indian quarry lies 700 to 800 feet west of the rock quarry (Fig. 1). It consists of a long north-south trench at the bottom of which is found the catlinite bed. Quarrying of the catlinite is accomplished by stripping the overlying quartzite from the eastern wall of the trench and throwing it back onto the quarry refuse heap. Between the modern rock quarry and

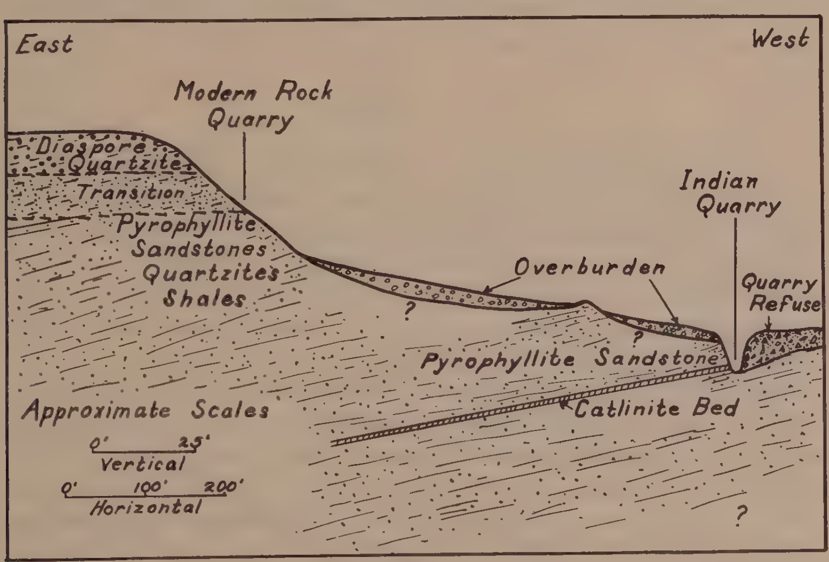


Fig. 1. Generalized cross-section through the Sioux quartzite at Pipestone, Minn. The quartzite at the top is cemented with diaspore. Below that, the cement is of diaspore and pyrophyllite (the transition zone). The lower portion of the outcrop is cemented with pyrophyllite. Quartz is an auxiliary cement throughout. See figures 2, 3 and 4.

the Indian quarry the quartzite is covered with glacial drift and slope wash, except for a distance of several feet where a quartzite bed outcrops.

The formation at this locality dips east at an angle of approximately 15°. Vertical joints are numerous and well developed. They belong to two main systems crossing each other at nearly right angles and striking N.E.–S.W. and N.W.–S.E.

Colors of the formation are variable, being predominantly shades of red and purple, but some of the thinner beds are nearly white. Along the joint planes the darker colored beds have often been bleached to a yellow or white color. So variable are the colors that in some instances it is impossible to trace an individual bed for more than a few feet horizontally.

The texture, excluding the fine-grained catlinite bed, varies from coarse sand to sandy shale. In the modern rock quarry the higher beds are the coarser while at lower horizons sandy shales are prevalent. Cross-bedding and ripple-marks may be seen in the sandy beds but they are not evident in the more shaly horizons. The sandy beds are of a somewhat uniform reddish purple color but greater variation occurs in the color of the shaly beds. All the beds are well cemented.

Samples were taken in the Indian quarry (a 6 foot section) and in the rock quarry (a 35 foot section). The vertical interval between samples was not uniform but averaged about 2 or 3 feet. The samples were numbered according to their relative stratigraphic position.

The Catlinite Bed

The catlinite bed is from 15 to 18 inches thick. It is massive but has partings parallel to the bedding which permit the stone to be broken into slabs 1 to 4 inches thick. No slaty cleavage is present. The pipestone is of a prevailing blood-red color but some of it has been replaced by a white or yellow-white mineral. The present study indicates that the catlinite, or red pipestone, is composed predominantly of sericite and that the light colored replacing mineral is pyrophyllite.

Pyrophyllite has replaced the top and bottom of the catlinite bed to a depth of 1 inch. Replacement has also taken place along the partings parallel to the bedding (see Fig. 6). Numerous lenses of pyrophyllite are found in the massive catlinite. These lie generally with their long axes parallel to the bedding. The lenses vary up to 4 inches in length. Pyrophyllite specks are scattered throughout the body of the catlinite and it would be difficult, on a polished surface, to find as much as a square inch of the red pipestone which did not contain these specks.

Minerals visible microscopically in the catlinite bed are sericite, pyrophyllite, diaspore, hematite (red), specularite and pyrite. Rutile is sus-

pected because of the titanium present (see chemical analysis), but it is probably indistinguishable from the hematite dust. The bulk of the diaspore (see Fig. 7) is found in the pyrophyllite replacements. The same is true of all the specularite and pyrite. Nearly all of the red hematite, a few grains of diaspore and the rutile (?) are found in the sericite.

The sericite was identified by x-ray and chemical methods. In thin section the red pipestone is seen to be composed of exceedingly fine grains with no definite orientation. These are too small to permit identification by optical methods. An x-ray diffraction pattern identified the bulk of the material as sericite since it showed only sericite (muscovite) lines.[7] Previous chemical analyses[8] of the red pipestone (all older than 1880) showed no potash, however, and this could not be reconciled with the x-ray determination. Therefore a new and more complete analysis was made. The specimen selected for the chemical analysis was blood red in color and nearly free from pyrophyllite specks. The analysis is given in Table 1. (Dr. R. B. Ellestad, analyst, Laboratory for Rock Analysis, University of Minnesota.)

TABLE 1

SiO_2	49.01	H_2O+	5.63
Al_2O_3	35.17	H_2O-	.24
Fe_2O_3	3.06	TiO_2	.44
FeO	none	Li_2O	.16
MgO	.23	Ignition, less total	
CaO	.05	H_2O	.24
Na_2O	.06		
K_2O	5.62		99.91

The chemical analysis supports the x-ray data in showing that the bulk of the material is sericite. If we assume that nearly all of it is sericite, the composition would be approximately $R_2O \cdot 5Al_2O_3 \cdot 12SiO_2 \cdot 5\ H_2O$. Although the potash content is lower than that of most muscovites, several secondary micas or sericites with about these molecular ratios have been reported.[9]

A microscopic study of the analyzed catlinite showed the presence of hematite, rutile (?), diaspore and pyrophyllite. A small amount of quartz is conceivably present but it has not been identified. The above minerals

[7] The writer is indebted to Dr. J. W. Gruner for the x-ray diffraction patterns and their interpretation.

[8] Winchell, N. H., *op. cit.*, p. 542, **1884.**

[9] Sericite from Carroll Driscoll Mine, Idaho. Shannon, E. V., *Bull.* **131,** *U. S. Nat. Mus.*, p. 367, 1926.

Secondary mica, Etta Mine, S. Dakota. Schwartz, G. M., and Leonnard, R. J.: *Am. Jour. Sci.*, vol. **11**, p. 262, 1926.

Sericite associated with pyrophyllite. Standard Mineral Co. Mine, N. Carolina. Stuckey, J. L., analyst: *Econ. Geol.*, vol. **20**, p. 457, 1925.

may make up as much as 10 per cent of the red stone, but if they were present in much larger amounts the diaspore would have increased the water content and the pyrophyllite and quartz (?) would have increased the silica content of the rock to figures larger than those given in the above chemical analysis.

The hematite, and probably the rutile, occur in the red pipestone as exceedingly fine grains or dust. While a thin section of catlinite is gray to tan in color, a section slightly thicker is dark red and nearly opaque because of the numerous hematite inclusions. Near the pyrophyllite replacements some of the catlinite is banded (see Fig. 6). This is undoubtedly due to the leaching of the hematite. The diaspore is present in small blades, up to .2 mm. in length, resembling those shown in figure 7 but not nearly as abundant as there shown. The pyrophyllite specks in the red pipestone may be seen in the dark band in figure 6.

The pyrophyllite was identified by *x*-ray and optical methods. The indices of refraction are: $\alpha=1.553$, $\beta=1.587$, $\gamma=1.598$; all $\pm.003$. The replacements consist of a number of small grains oriented in all directions. Because of its high birefringence these differently oriented pyrophyllite grains are outlined even in plain light (Fig. 7). In most instances the pyrophyllite replacements are light in color and are composed of pyrophyllite with a few diaspore grains. All of the lenses described above, and a few of the bedding plane replacements, are however of a dark reddish-purple color. This color is due to the large amounts of red diaspore and specularite which are present. A small amount of pyrite in little striated crystals accompanies the diaspore and specularite. The specularite crystals measure up to .25 mm. in size and are easily seen in the hand specimen because of the brilliant reflection from their crystal faces. The specularite is probably due to recrystallization of the red hematite present in the catlinite. Had there been ferrous iron in the catlinite it is possible that magnetite would have formed.

The specific gravity of a specimen of catlinite is 2.746 while that of pyrophyllite is 2.754, as determined by heavy liquids and the Westphal balance. Both the catlinite and the pyrophyllite appear to have little pore space so that the correct densities are probably not far above those given.

The catlinite has a hardness which is somewhat variable but averages about 2.5. The pyrophyllite is approximately 2.0 in hardness. It is commonly believed that catlinite is soft when quarried and hardens upon being dried. A foot or so of water is usually present on the bottom of the Indian quarry so that the lower portion of the catlinite bed is immersed. Visitors to the quarry frequently encounter difficulty in attempting to secure pipestone that lies under the water. The writer was un-

able to find any difference in hardness between the wet pipestone at the bottom of the quarry and the slabs which had been thrown back on the refuse heap. Samples taken from the bottom of the quarry showed no appreciable difference in hardness six months later. To test the hardness further a piece was sawed in two, one half was saved for comparison and the other half was placed in an oven. After heating 100 hours at 200°C. there was no discernible difference in hardness between the two pieces. The analyzed sample, which showed 0.24 per cent of absorbed water, had been dried three months at room temperature.

Quartzite Beds

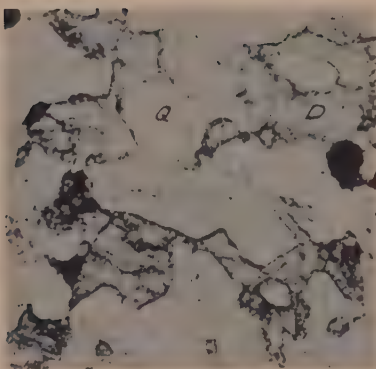
The quartzite beds which overlie the catlinite at Pipestone may be divided into three types on the basis of the cementing material: first, those which are cemented by diaspore and quartz; second, those cemented by diaspore, pyrophyllite and quartz; third, those cemented by pyrophyllite and quartz.

The diaspore quartzites (Figs. 1 and 2) have been described by the writer in a previous paper.[10] They do not extend deeper than 5 or 6 feet from the surface at this locality. At their lower surface they give way to the diaspore-pyrophyllite quartzites in which grains of diaspore appear to be developing in a groundmass of pyrophyllite (Fig. 3). This is the transition zone of figure 1 and is almost 10 feet in thickness. Below this zone there is no diaspore, except that in the catlinite bed. Figures 1, 4, and 5 illustrate the relations of the beds below the diaspore. The quartzite beds immediately below the impervious catlinite layer are pyrophyllitized in the same manner as those immediately above.

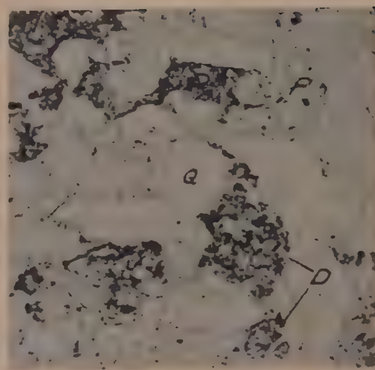
Eighteen thin sections of the various types of quartzite were examined. Two were from the diaspore facies, four from the diaspore-pyrophyllite facies and twelve from the pyrophyllite beds. In addition, several samples from the pyrophyllite zones were crushed and subjected to a bromoform separation. The heavy minerals in each case were iron oxides and zircon. No diaspore was observed. The thin sections showed the presence of quartz, chalcedony, pyrophyllite, diaspore, iron oxides, rutile and zircon in the quartzite beds.

The pyrophyllite in the quartzite beds was identified optically, but as a check several samples of both the red and white varieties of quartzite were crushed, and screened to remove the coarse grains of quartz. The fine grained fraction was *x*-rayed and was found to contain pyrophyllite and quartz.

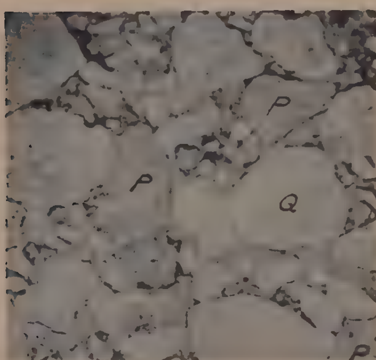
[10] Berg, Ernest L., An occurrence of diaspore in quartzite: *Am. Mineral.*, vol. **22**, pp. 997–999, 1937.



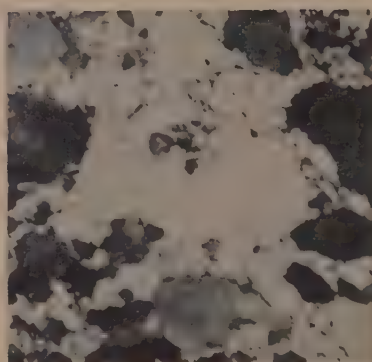
2



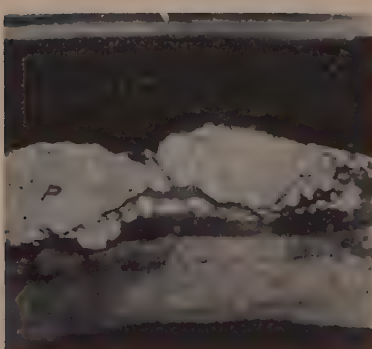
3



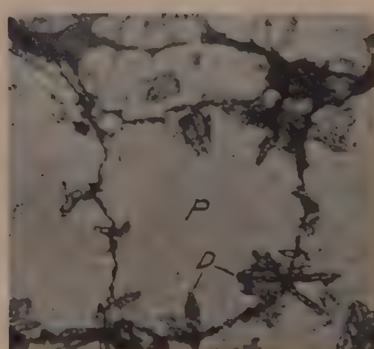
4



5



6



7

PHOTOMICROGRAPHS OF THIN SECTIONS.

FIG. 2. (Upper left) Quartzite cemented by diaspore (D). Plain light (×30)

3. (Upper right) Quartzite cemented by pyrophyllite (P) in which diaspore is beginning to form. Plain light (×30)

4. (Middle left) Quartzite cemented by pyrophyllite. Note the grains of pyrophyllite which are apparently replacements of other minerals. Plain light (×30)

5. (Middle right) Same as figure 4. Crossed Nicols (×30)

6. (Lower left) Pyrophyllite replacement parallel to the bedding of the pipestone. Plain light (×5)

7. (Lower right) Pyrophyllite replacement in pipestone containing minute seams (iron-stained) along which diaspore crystals have formed. Plain light (×60)

In most of the sandy shale horizons the individual quartz grains are completely surrounded by pyrophyllite. Some of the grains are partially or wholly replaced by this mineral. A thin section of the quartzite which immediately overlies the catlinite bed shows that almost all of the quartz grains within half an inch of the catlinite bed have been replaced by pyrophyllite. The identity of the replaced mineral can be determined because a few of the grains have not been entirely replaced. These show a small remnant of quartz. Furthermore, one would not expect to find many grains of a mineral other than quartz in a sandstone. Figures 4 and 5 show several replaced grains in a bed several feet above the catlinite. In this material no remnants of the original material can be observed. It may have been feldspar, quartz or still some other mineral.

All of the thin sections showed that the quartzite formation is well cemented at this locality. Not all of the rocks are quartzites, however, since those cemented by pyrophyllite are too soft to be so classified. Also, some beds are too fine textured to be called quartzite, but the term quartzite may be used to distinguish the rest of the formation from the catlinite bed.

Certainly the most striking feature of the quartzite beds is the relation of the diaspore to the present surface. At the bottom of the section the beds are cemented with pyrophyllite but, as one approaches the surface, small grains of diaspore appear and finally at the surface the pyrophyllite gives way entirely to diaspore.

Probable History of Deposits

Summarizing the foregoing statements it is seen that the Sioux quartzite at this locality is a clastic sedimentary deposit of Algonkian age varying in texture from coarse sand to very fine shale. The formation has been subjected to static metamorphism but not to intense dynamic action since the catlinite bed shows no slaty cleavage and there is no parallelism of grain.

At some time during the consolidation or metamorphism of the formation, the sericite which makes up the bulk of the catlinite, was formed from the minerals of the original shale. Large amounts of seri-

cite are not unusual in metamorphosed shales.[11] Dale[12] has shown that the red slates of Vermont, New York and elsewhere are largely composed of sericite and quartz. He quotes Hutchins[13] as follows:

> . . . the fine mixture of biotite, muscovite, kaolin, the minutest waste of feldspar, and in less degree of quartz . . . under the joint action of pressure, warmth, and mineral solutions, gives rise to various decompositions and recombinations which result, among other things, in the formation of new mica.

However, the absence of quartz in discernible amounts is noteworthy. If the original shale contained only moderate amounts of quartz it might have been consumed in the formation of the sericite, or it might have been removed in solution, or entered into the composition of pyrophyllite when that mineral replaced portions of the sericite. It is possible, of course, that the catlinite was originally a volcanic ash or bentonite bed containing little or no quartz. Twenhofel reports three bentonites which contain only from 3 to 7 per cent of quartz.[14] Wherry[15] reports a deposit which shows no quartz. But analyses[16] of these bentonites show that they are low in alumina and potash, and high in lime and magnesia.

Subsequent to the formation of the sericite, which makes up the catlinite bed, replacement by pyrophyllite began. Very likely the replacement of the quartzite beds took place at the same time. The pyrophyllitization of the quartzite beds implies that they were originally somewhat arkosic and argillaceous. Potash feldspar, kaolin and sericite are commonly present in arkoses and shaly sediments. These minerals are known to yield pyrophyllite under hydrothermal conditions both in the laboratory and in the field.

Buddington[17] suggests the following reaction for the alteration of sericite to pyrophyllite:

$$\underset{\text{(sericite)}}{2KH_2Al_3Si_3O_{12}} + 6SiO_2 + H_2O \rightleftarrows \underset{\text{(pyrophyllite)}}{6HAlSi_2O_6} + K_2SiO_3$$

It was suggested above that any quartz originally in the catlinite

[11] Van Hise, C. R., A treatise on metamorphism: *U. S. Geol. Survey, Monograph* **47,** 1904.

[12] Dale, T. Nelson, Slate deposits and slate industry of the United States: *U. S. Geol Survey, Bull.* **275,** 1906.

[13] Hutchins, W. M., Clays, shales and slates: *Geol. Mag.*, vol. **7,** p. 317.

[14] Twenhofel, W. H., *Treatise on sedimentation: Second Edition*, p. 270, **1932.**

[15] Wherry, E. T., Clay derived from volcanic dust in the Pierre of S. Dakota: *Jour. Wash. Acad. Sci.*, vol. **7,** pp. 576–583, 1917.

[16] Twenhofel, W. H., *op. cit.*, p. 268.

[17] Buddington, A. F., Pyrophyllitization, pinitization, and silicification of rocks around Conception Bay, Newfoundland: *Jour. Geol.*, vol. **24,** p. 147, 1916.

might possibly have gone into the formation of pyrophyllite according to Buddington's equation. The numerous specks of pyrophyllite found throughout the catlinite could then be interpreted as replacements of quartz. At any rate the equation shows that silica and water must have been added to the sericite, and potassium silicate removed.

Since the formation of pyrophyllite is usually attributed to hydrothermal alteration,[18] and since specularite is commonly thought of as a high temperature mineral,[19] the presence of these minerals implies at least some thermal metamorphism in the rocks of the Pipestone region. There are no known intrusives in the vicinity that are younger than the quartzite. There is, however, some evidence that pyrophyllitization is not limited to the Pipestone area. At New Ulm, Minnesota, 100 miles to the east, pyrophyllite and diaspore are present in the Sioux quartzite. Diaspore, which is associated with the pyrophyllite at Pipestone, is found widely distributed in surface samples of the Sioux quartzite.[20] Pyrophyllitization may, then, be general throughout the formation. Statements as to the temperatures and types of metamorphism should be postponed until more is known of other portions of the formation.

The occurrence of diaspore near the surface strongly *suggests* that pyrophyllite may be unstable under certain environmental conditions, breaking down into diaspore and quartz according to the reaction:

$$\underset{\text{(pyrophyllite)}}{H_2Al_2Si_4O_{12}} \rightarrow \underset{\text{(diaspore)}}{Al_2O_3 \cdot H_2O} + \underset{\text{(quartz)}}{4SiO_2}$$

The silica need not be removed but could be deposited as a secondary growth on the original quartz grains. It must be remembered, however, that diaspore is present in the catlinite bed. Here, too, it is associated chiefly with pyrophyllite and appears to have formed along minute seams (Fig. 7). Nevertheless the catlinite layer is somewhat removed from the surface so that the relation of the diaspore in the catlinite bed to the diaspore in the quartzite is not clear.

Summary and Conclusions

1. The Sioux quartzite at Pipestone, Minnesota, was originally an arkosic sandstone with interbedded argillaceous strata, some of which were very fine grained. The catlinite bed represents the finest of these argillaceous layers.

[18] Buddington, A. F., *op. cit.*

Stuckey, J. L., *Econ. Geol.*, vol. **20**, p. 457, 1925.

Vhay, J. S., (abstract) *Am. Mineral.*, vol. **22**, no. 12, part 2, p. 15, 1937.

[19] Rogers, A. F., Origin of copper ores of the "red bed" type: *Econ. Geol.*, vol. **11**, p. 378, 1916.

[20] Berg, Ernest L., *op. cit.*

2. During consolidation or metamorphism most of the catlinite bed was converted into secondary mica (sericite) and small amounts of free quartz.

3. The catlinite shows neither secondary cleavage nor any evidence of recrystallization under dynamic metamorphism. Metamorphic agencies, probably hydrothermal, produced pyrophyllitization of the quartzite beds and portions of the catlinite bed. Pyrophyllite probably replaced most of the quartz originally present in the catlinite. Specularite was developed locally in the pyrophyllitized pipestone.

4. Near the surface of the outcrop and in the catlinite bed, exposed in the quarry, the pyrophyllite *appears* to have broken down yielding diaspore and quartz.

Acknowledgments

The writer is indebted to Dr. J. W. Gruner for x-ray identification of several minerals. Dr. R. B. Ellestad and Mr. Lynn Gardiner gave invaluable assistance in the preparation of this paper.

URANINITE AND ASSOCIATED MINERALS FROM HADDAM NECK, CONNECTICUT

Earl Ingerson,
Geophysical Laboratory, Carnegie Institution of Washington.

Introduction

Many fine uraninite crystals have been found in the Connecticut pegmatites in Glastonbury, Portland, and Branchville. Schairer[1] mentions torbernite from Haddam Neck, but does not record uraninite from that locality.

In the fall of 1932 a box of minerals from a pit that had been opened a short time before at the Rock Landing Quarry, Haddam Neck, was brought to Yale University for identification. The material consisted mostly of smoky quartz and black tourmaline, but there were two remarkable specimens of uraninite, and one crystal and several fragments

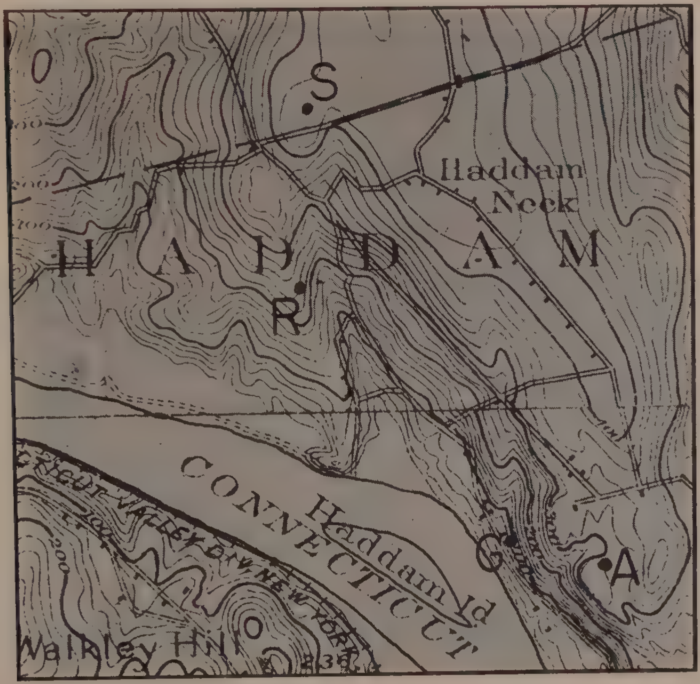


Fig. 1. Index map. S—Swanson Quarry; R—Rock Landing Quarry; G—Gillette Quarry; A—pit from which a number of gem aquamarines have been taken.

[1] Schairer, J. F., The minerals of Connecticut: *Connecticut State Geol. and Nat. Hist. Survey, Bull.* no. **51**, p. 69, 1931.

of columbite. There were also numerous patches and crystals of autunite and torbernite on the other minerals.

Locality

There are several quarries at Haddam Neck and they are somewhat different in the minerals that they yield. The index map, figure 1, shows the locations of these quarries. Lithium minerals have been found in all of these quarries except the Rock Landing Quarry, and it is the only one from which uranium minerals have been reported. Gem tourmaline and beryl crystals have been found in the other quarries, but only black tourmaline and opaque beryl are known at Rock Landing. Triplite is found only at the Swanson Quarry; microlite, red fluorite, and cookeite, only at the Gillette Quarry.

Minerals from the Rock Landing Quarry

1. Uraninite

The uraninite was found in a single pocket, although its alteration products are rather widely scattered in the quarry. In the material studied there were two pieces of fairly pure uraninite, one piece consisting of an intimate mixture of uraninite and columbite, and numerous small fragments of uraninite. The largest piece is $6\times7\times10$ cm. and weighs 1,200 grams. The next largest piece is $2\times3\times4$ cm. and weighs 89 grams. In the large piece there is one octahedral crystal 12 mm. in diameter (Fig. 2), and there are numerous smaller but better formed octa-

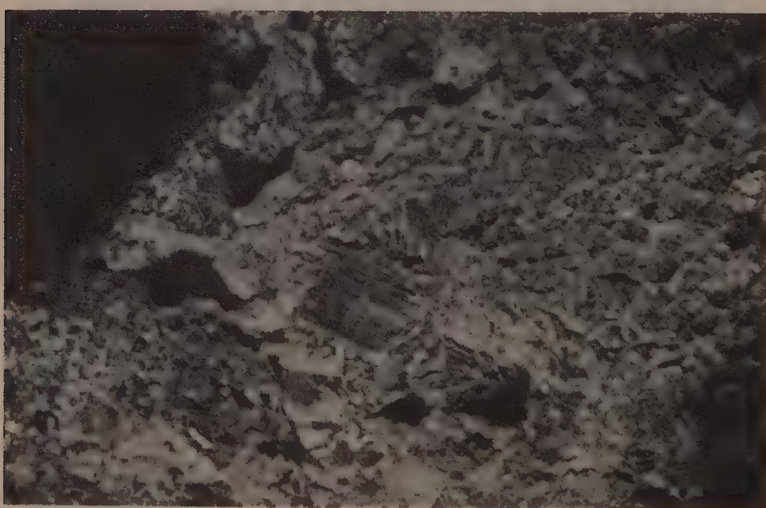


Fig. 2. Large uraninite crystal on a face of the largest specimen. $\times2$.

hedra from 0.5 to 3.0 mm. in diameter. Some of the crystals show cubic and dodecahedral faces.

Identification: (a) The association and crystal form indicated that the mineral was probably uraninite.

(b) *Etch tests:* The smaller specimen was polished on one side (Fig. 5A) and the tests given by Short[2] were applied. The material is completely isotropic, and gives no reaction with HNO_3, HCl, KCN, KOH, $HgCl_2$, or aqua regia. It is slowly darkened by $FeCl_3$. Streak, greenish black; hardness, G−.

(c) *Microchemical test for uranium:* The standard sodium acetate test[3] for uranium was applied and figure 3 shows the tetrahedra of sodium uranyl acetate that were formed.

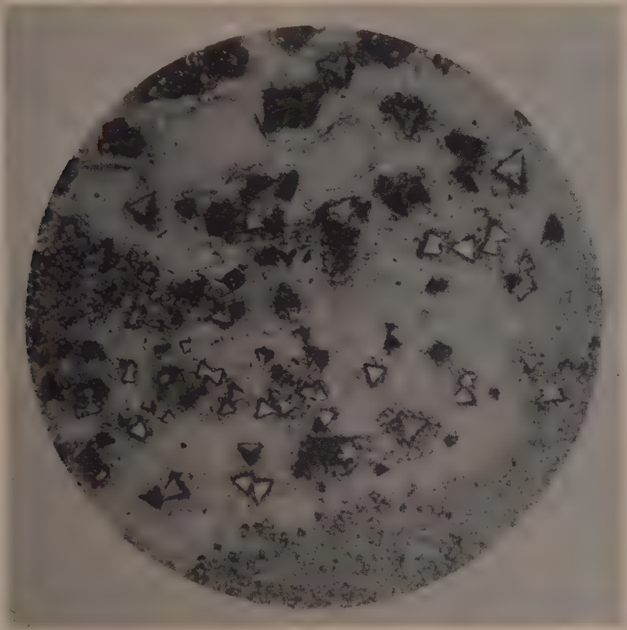


FIG. 3. Photomicrograph of the microchemical test for uranium; tetrahedra of sodium uranyl acetate.

(d) *Radioactivity:* To prove that the material is radioactive a star cut out of sheet lead was placed between the large piece and a commercial orthochromatic photographic film and the set-up was left in a dark room

[2] Short, M. N., Microscopic Determination of the Ore Minerals, *U. S. Geol. Survey, Bull.* no. **825**, p. 102.

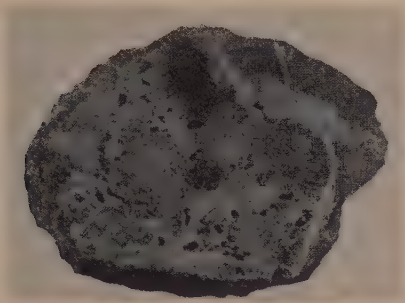
[3] Chamot, E. M., and Mason, C. W., *Handbook of Chemical Microscopy*, vol. **2**, pp. 52–58 and 225.

for nine days. The film was developed for seven minutes in stock pyro film developer. Figure 4 shows the result. The smaller piece took a recognizable picture of itself on x-ray film in eleven days. Figure 5B shows this picture and Fig. 5A is a photograph of the specimen.

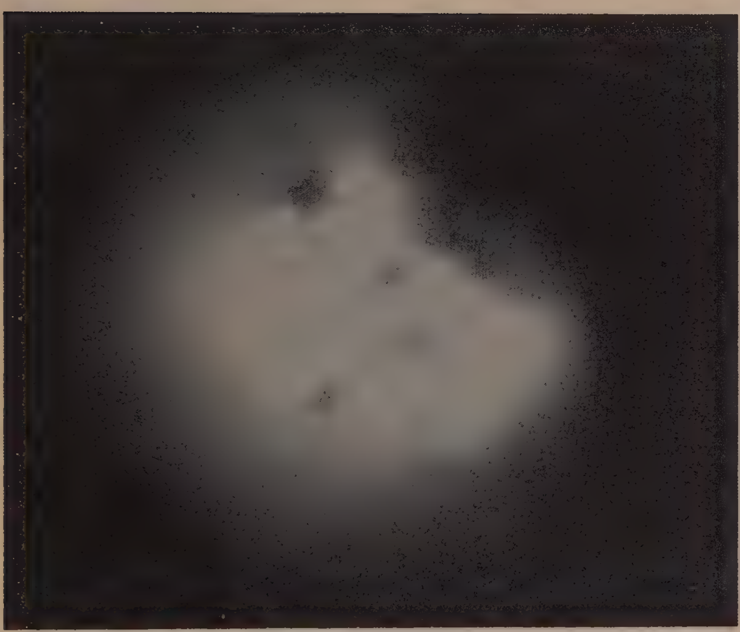
FIG. 4. Shadow of a star-shaped piece of lead made by placing the lead between the large piece of uraninite and a photographic plate.

(e) *X-ray:* Mr. C. J. Ksanda, of this Laboratory, powdered a fragment of a small crystal and made an x-ray powder photograph from it. The crystal was carefully selected under the microscope. It was of uniform color and showed several well-developed dodecahedral faces. The photograph showed only lines belonging to a cubic pattern of face-centered lattice type with $a_0 = 5.454 \pm .006$ Å (see the table for the planar spacings and relative intensities of these lines). The density calculated from this value of a_0 is 10.98, on the assumption that the unit cell contains four "molecules" of UO_2.

The source of monochromatic x-radiation was an ion gas discharge metal tube. CuK-radiation filtered with Ni foil was used. The precision circular camera used has a radius of 57.22 mm. and was calibrated with NaCl.



A



B

FIG. 5. A. The second largest piece of uraninite after polishing. Natural size. B. Effect of the face shown in A on a photographic plate. Natural size. Note the dark patches where feldspar occurs on the specimen.

TABLE 1. PLANAR SPACINGS AND RELATIVE INTENSITIES OF THE X-RAY DIFFRACTION LINES OF URANINITE FROM POWDER SPECTRA WITH FILTERED CuK-RADIATION

Line	d/n	Intensity observed*	Indices hkl	a_0
1	3.147	10	111	5.451
2	2.728	6	200	5.456
3	1.928	8	220	5.453
4	1.645	8	113	5.455
5	1.575	2	222	5.456
6	1.364	2	400	5.456
7	1.251	4	133	5.453
8	1.220	4	204	5.456
9	1.113	5	224	5.453
10	1.050	5	115, 333	5.456
11	.964	1	440	5.453
12	.922	5	135	5.455
13	.909	4	244, 600	5.454
14	.862	4	260	5.453
15	.832	3	335	5.455
16	.822	3	226	5.453

* Estimated intensities of the lines are based on a scale of ten, 10 being the strongest line.

Age determination: Hecht and Kroupa[4] made a micro-analysis of part of one of the uraninite crystals from the larger of these two specimens from the Rock Landing Quarry and obtained

$$\frac{\mathrm{Pb}}{\mathrm{U}+0.36\ \mathrm{Th}}=0.040$$

This gives an age of between 280 and 290 million years, which is in good agreement with results of analyses of radioactive minerals from other Connecticut pegmatites. It places the time of formation as late Devonian.

2. GUMMITE

The uraninite crystals of the two largest specimens are surrounded by yellow waxy gummite which penetrates the feldspar thoroughly, but is excluded from the individual uraninite crystals except along small and rare cracks.

[4] Hecht, F., and Kroupa, E., Die Bedeutung der quantitativen Mikroanalyse radioaktiven Mineralien für die geologische Zeitmessung: *Zeit. anal. Chemie*, vol. **106**, pp. 83–103, 1936. This analysis is also given on p. 61 of the May, 1936, report of the committee of the National Research Council on geologic time, and is mentioned in the 1937 report.

3. Autunite

This mineral is present on several of the fragments of quartz in the material that was brought in, as well as on the larger pieces of uraninite. It occurs for the most part as almost micaceous coatings that apparently were fillings of seams in and between the other minerals. Much of the autunite is in the form of feathery flat needles up to 3 mm. in length, and isolated pseudo-tetragonal plates are rare. Autunite is widely distributed in the new pit at the quarry and it is fairly easy to find small patches of it either on the quarry wall or on loose fragments. Uraninite may be almost as widely distributed, but it is much more difficult to find owing to the abundance of dark smoky quartz and black tourmaline in the quarry.

4. Torbernite

Torbernite occurs as apple green crystals much like those of the autunite in size and shape, and also as seam fillings. Its distribution in the quarry is also much like that of autunite, but it is much less abundant. Both of these minerals were studied under the microscope. They are both negative, practically uniaxial, and their indices fall within the range commonly given.

Fig. 6. Crystal of columbite from the Rock Landing Quarry.

5. Columbite

Columbite is also difficult to recognize in the quarry because of the associated minerals. In most of the hand specimens the columbite is readily distinguished from uraninite by its crystal form and its characteristic iridescent tarnish. The best crystal in the material studied is shown in figure 6. It is a prismatic crystal that measures $1.4 \times 1.4 \times 4.5$ cm. The following angles were measured with an ordinary contact goniometer:

$$110 \wedge 1\bar{1}0 = 80°$$
$$110 \wedge 100 = 40°$$

6. Other Minerals

The other minerals to be found at the Rock Landing Quarry are common pegmatite minerals that do not warrant special descriptions—smoky quartz, muscovite, potassium feldspar (perthite), black tourmaline, apatite, and large beryl crystals. There are also many small crystals of chalcopyrite that may have served as the source of copper for the torbernite, or may have been formed from copper-bearing solutions at the same time the torbernite was formed. The quartz and feldspar are in places intergrown as graphic granite. The black tourmaline is in part well crystallized, the rounded, well-terminated crystals commonly being as much as 2 cm. in diameter and 5 cm. long. On many of these crystals there are pyrite cubes up to 2 mm. on a side.

THE FORMATION OF IDDINGSITE

A. B. Edwards,
University of Melbourne, Australia.

Introduction

In their comprehensive paper on the origin, occurrence, composition and physical properties of the mineral iddingsite, $MgO \cdot Fe_2O_3 \cdot 3SiO_2 \cdot 4H_2O$, Ross and Shannon (5) concluded that iddingsite is not a product of weathering, but is a deuteric mineral derived during the final cooling of the lavas in which it occurs by a reaction between gases, water, and olivine, under oxidizing conditions. They claimed further, that "the magma must have come to rest before iddingsite formed, for though it is a very brittle mineral, it is never fractured or distorted by flow."

Late Formation of Iddingsite

This claim is confirmed by the manner of occurrence of iddingsite in basalts from two Victorian localities, Lady Julia Percy Island (4), and the Colac District (6).

At Lady Julia Percy Island, near Portland, a bed of "boulder tuff," about 30 feet thick, occurs at the base of six flows of olivine basalt, the upper five of which contain abundant iddingsitized olivine.

The boulders of the tuff consist of a core of dense crystalline basalt grading outwards into a margin of tachylyte. In thin section the tachylyte is seen to consist of a light greenish-brown glass, in which are set numerous small, and a few large, crystals of olivine, and scattered laths of basic plagioclase (Ab_{35}). The olivine crystals, especially the larger ones, are much corroded, but the olivine is fresh and unaltered.

Towards the interiors of the boulders, where crystallization has been more or less complete, the olivine crystals are rimmed with iddingsite. The iddingsite is clearly not a weathering product, nor could it have formed previous to the extrusion of the lava boulders. Rapid escape of gases from the magma, and sudden chilling of the surface of the boulders, preserved the olivine unaltered in the tachylyte margins. This freezing of the rim, however, retained some water within the central part of the boulders, and also slowed the rate of cooling of the interior, making iddingsitization possible.

Thin selvages of tachylyte frequently form a "skin" to basalt flows in the Colac district (6). The tachylytes are identical with that just described from Lady Julia Percy Island. They consist of corroded phenocrysts of unaltered olivine, and scattered laths of labradorite, set in a light brown, or sometimes black, glass, Where, however, the tachylyte

begins to grade into a more crystalline basalt, iddingsite appears irregularly in the cracks, and at the rims of the olivine crystals. In certain of the crystalline lavas it is a common mineral constituent. Here again, the iddingsite was formed during the consolidation of the lava, but subsequent to its extrusion, and after it had come to rest.

In both instances it is clear that the magma was at a relatively high temperature when extruded, since pyroxene, although a common phenocryst constituent of the crystalline lavas, had scarcely commenced to crystallize out of the tachylyte.

Earlier Formation of Iddingsite

Where, however, the magma had cooled to a somewhat lower temperature before extrusion, the iddingsite seems to have formed at a slightly earlier stage, viz., during the actual process of extrusion, and before the lava came to rest.

Thus at the summits of many of the centres of eruption of the Newer Volcanic basalts of Victoria, highly glassy, but not tachylytic, rocks are found. These consist of numerous fine laths of plagioclase, with phenocrysts of augite and iddingsitized olivine, set in a dense dark glassy base. The darkness of the glass is due to the presence in it of innumerable globulites of iron oxide. These rocks are, moreover, finely, but highly

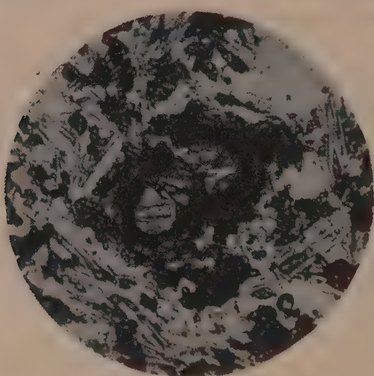


Fig. 1

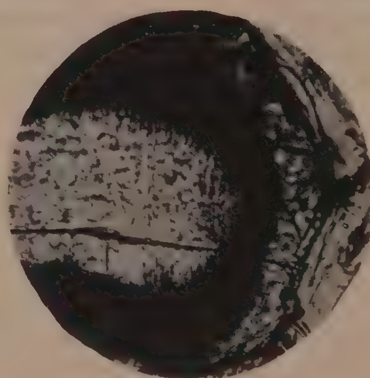


Fig. 2

Fig. 1. Iddingsite round a core of embayed olivine, with a margin of later olivine. The later olivine tends to have an idiomorphic outline, which is in part masked by a number of small grains of augite, which are moulded on the olivine. Ordinary light, ×18.

Fig. 2. Magnified view of the olivine crystal shown in figure 1. The junction between the iddingsite and the core olivine is fibrous, denoting a reaction, while the junction with the outer olivine is sharp, suggesting a cessation of iddingsite-forming conditions and a return to olivine precipitation. The idiomorphic outline of the outer olivine is clearly shown. The contraction cracks in the olivine have formed subsequent to development of iddingsite. Ordinary light, ×146.

vesicular, testifying to the presence and easy escape of an abundance of gases.

Two observations indicate that the iddingsite was formed before consolidation was complete:— (1) rims of augite surrounding iddingsitized olivine crystals, and (2) zones of fresh olivine, often idiomorphic in outline, enclosing embayed and iddingsitized olivine phenocrysts (Figs. 1 and 2).

Ross and Shannon (5) described "large phenocrysts (of olivine) completely altered to iddingsite while small groundmass olivines of a later generation show little alteration," and "cores of iddingsite surrounded by fresh olivine." They concluded from this "that the alteration was partly dependent upon the zonal variations in the original olivine," iddingsite only forming in the presence of olivine of a certain limited chemical composition.

In the Victorian basalts one finds, not infrequently, crystals such as that portrayed in figure 1, in which is shown an irregular shaped core of olivine partially replaced by a concentric rim of iddingsite, while outside this is a further rim of olivine with a partially idiomorphic outline (Fig. 2), more or less surrounded by grains of augite. The shape of the iddingsite zone is such as to indicate that it does not replace a well defined zone of olivine of particular composition, but is concentric with the shape of a corroded olivine crystal. Moreover, as figure 2 shows, the junction between the core olivine and the iddingsite is fibrous, denoting reaction, while the junction between the iddingsite and the outer olivine is sharp, pointing to a sudden cessation of reaction, and a reversal to olivine precipitation. This is a common feature in those Victorian basalts which contain iddingsite.

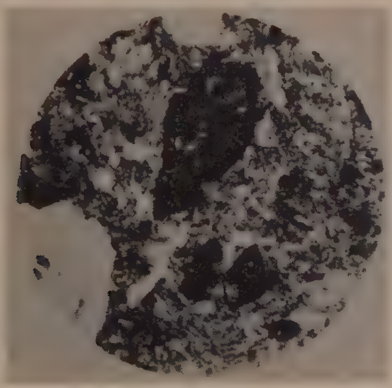


FIG. 3. Pseudomorph of iron ore after iddingsite, itself a pseudomorph after olivine. Ordinary light, ×26

The formation of iddingsite is a process of oxidation and hydration:

$$2(Mg,Fe)O \cdot SiO_2 \rightarrow MgO \cdot Fe_2O_3 \cdot 3SiO_2 \cdot 4H_2O$$

In some instances, however, the action has gone further, and a rim of iron oxide has formed on the outer margin of the iddingsite. Eventually all of the original olivine vanishes, and the iddingsite which had formed a rim about it is completely replaced by magnetite (Fig. 3):

$$MgO \cdot Fe_2O_3 \cdot 3SiO_2 \cdot 4H_2O \rightarrow FeO \cdot Fe_2O_3$$

Goodchild (3) states that the reaction:

$$H_2O + 2FeO \rightleftharpoons H_2 + Fe_2O_3$$

is a delicately balanced, isothermal reaction. Since water vapour does not dissociate at atmospheric pressures, even at 2,000°C. it is probable that such a system is stable under the conditions which exist during and immediately prior to extrusion. In the presence of sufficient water vapour, and with the removal of Fe_2O_3 from further reaction, the process would develop in the left to right direction. With escape of water vapour, this action would come to a stop, and the right to left reaction might set in. Hence it is suggested that the formation of iddingsite would cease soon after the extrusion of the lava, since at that period the water vapour would have escaped.

It seems, therefore, that in such magmas as reach a sufficiently low temperature before extrusion, the iddingsite would develop during the rising of the magma towards extrusion, when as it neared, and passed through the vent, the mineralisers would be most concentrated. Aurousseau (1) has come to somewhat similar conclusions from his examination of the nepheline basanite of Sandy Bay, Tasmania. He writes . . . "these facts suggest that the iddingsite is the result of oxidizing processes that acted rapidly on the olivine, during the liberation of copious volatile phases at the time of extrusion."

It is essential for the formation of iddingsite that the magma should not only be rich in water vapour, but that it should have differentiated in such a manner as to give rise to an iron-rich final fluid. It is characteristic of the Victorian basalts that those rich in iddingsite possess such iron-rich glasses, while in those from which the iron oxides had completely crystallized, no iddingsite formed, although chemically the rocks might be identical (2). In the "non-iddingsitized" basalts the olivine remains fresh, or is altered to serpentine.

The presence or absence of iron oxide globules in the glass is probably controlled by the temperature of extrusion, a low temperature of extrusion providing conditions favouring the completer crystallization of the

glass, and so reducing the amount of iron oxide in the volatiles. Only rarely has iddingsite been observed accompanying iron ores which have crystallized at an early stage.

Conclusions

It is concluded, therefore, that iddingsite forms during, or after, extrusion, according to the temperature of the magma at the time of extrusion; and that if the magma has cooled sufficiently before extrusion not enough iron oxides are left in the residual volatiles for iddingsite to be formed.

References

1. Aurousseau, M., *Proc. Lin. Soc. N. S. W.*, vol. **51**, 4, p. 617, 1926.
2. Edwards, A. B., Tertiary volcanic rocks of central Victoria: Abstract No. 1322, *Geol. Soc. London*, Feb. 10th, 1937.
3. Goodchild, W. H., *Mining Magazine*, p. 26, 1918.
4. McIvor, E., *Proc. Roy. Soc. Victoria*, n.s., vol. **49**, 2, p. 344, 1937.
5. Ross, C. S., and Shannon, E. V., The origin, occurrence, composition and physical properties of the mineral iddingsite: *Proc. U. S. Nat. Mus.*, vol. **67**, pp. 1–19, 1925.
6. Skeats, E. W., and James, A. V. G.: *Proc. Roy. Soc. Victoria*, n.s., vol. **49**, 2, pp. 250–51, 1937.

NOTES AND NEWS

THE OSSEO, CANADA, METEORITE

JOHN PUTNAM MARBLE,
Chevy Chase, Maryland.

HISTORY

This iron meterorite was received by the United States National Museum in May, 1935, from Mr. Frank Johnston, Cane P.O., Ontario, Canada. Mr. Johnston writes:[1] "This meterorite was found three miles from the village of Osseo, Temiskaming District, Northern Ontario, and was lying on a flat rocky knoll. This meterorite was on a solid flat rock, which I have examined closely, but cannot find the slightest mark where it fell." "This meteorite was found in June, 1931, was not a witnessed fall, and was found by accident by a settler doing a little prospecting on his lot." It is further reported that it was at first taken to be a mass of metallic silver.

It is listed at the United States National Museum, Washington, D.C., under No. 925. Several slices have been cut from the main mass, and some have been distributed to other collections, as follows:

Stuart Perry, Adrian, Michigan	750 grams.
Colorado Museum of Natural History, Denver, Colo.	775 grams.
H. H. Niniger, Denver, Colorado	140 grams.
Kyancutta Museum, Kyancutta, Australia	290 grams.
Geological Survey of India, Calcutta, India	505 grams

DESCRIPTION

The iron as received was an irregularly oblong mass, with one rather flat face, the opposite face somewhat domed. The most striking features are the marked flutings that characterize one side of the meterorite. These are straight-edged grooves, varying from very shallow, as in the left-hand side of the photograph, figure 1, to over 2 cm. deep, in the right center. They average about 2 cm. in width. These grooves are not in parallel alignment, but are inclined somewhat to each other. If they are due to scouring action during the passage of the mass through the atmosphere, some variation in direction during flight might be suggested.

The meterorite is covered by the usual oxidized crust, which is not of any great depth. The top and bottom have the usual pitted surfaces. The original weight of the mass as received was 46.3 kilograms.

A polished slice, etched with dilute nitric acid in alcohol, shows this

[1] Private Communication, in files of U. S. National Museum.

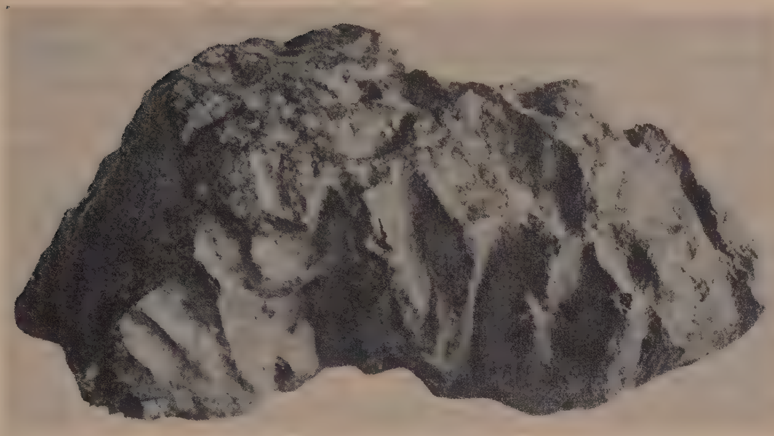


FIG. 1. The Osseo, Canada, meteorite. Side view. $\frac{1}{3}$ natural size.

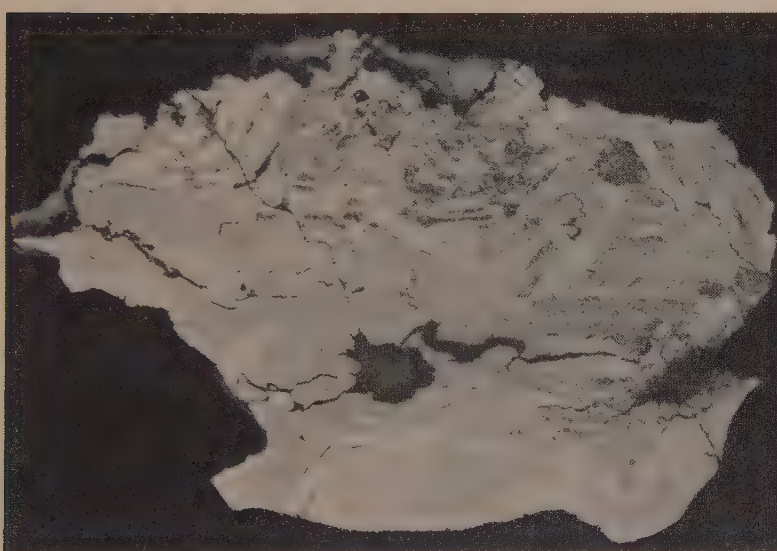


FIG. 2. Osseo, Canada, Meteorite. Polished, etched slice, 4/5 natural size.

mass to be a very coarse octahedrite. Random measurements made on several grains show the longer dimensions to vary between 9 and 26 mm., the shorter between 1 and 8 mm. It is a kamacitic iron, carrying a few troilite nodules and rather rare schreibersite. In places it shows cracking along the kamacite individuals to give a somewhat brecciated appearance. The Neumann lines are bent in places, which is a somewhat

unusual feature, and may perhaps indicate distortion during the flight, as may also the arrangement of the exterior grooves, as noted above.

On the slice studied, one of the troilite nodules is surrounded by schreibersite, and there is also considerable schreibersite having no connection with the troilite. The two main troilite nodules are connected with a narrow band of troilite, which varies in width, and seems to follow the kamacite boundaries in a general fashion. The schreibersite seems oriented irregularly in the same network as the kamacite. One or two small areas of plessite may be seen, and some oxidized troilite in thin bands, more or less between kamacite grains. There seem to be some areas of granular iron along the outside boundary of the slice. No lawrencite is visible. See figure 2.

Analysis

A portion weighing about 25 grams was taken for analysis. No troilite and very little schreibersite were visible on its surface. The thin oxidized crust was removed by grinding. The results are shown in Table 1.

Table 1. Chemical Analysis of the Osseo, Canada, Meteorite

C	0.007%	Co	0.11
Pt, etc.	0.02	Ni	6.51
S	0.00	Fe	92.89
P	0.50		———
Cu	0.10	Total	100.137%

J. P. Marble, *Analyst.*

No striking peculiarities appear in this analysis. The ratio of iron to nickel is well within the average range of coarse octahedrites. A trace of chromite may be included in the small fraction reported as platinum, as this was too small for an accurate separation into further constituents. The absence of troilite from the analyzed portion is borne out by the absence of sulphur. The low phosphorus content shows only minor amounts of schreibersite. Evidently the troilite and schreibersite are irregularly distributed. Except for the carbon and platinum portions, the entire sample was soluble in warm dilute nitric acid.

Comments

The absence of any marks on the ground underlying the meteorite when found may perhaps be accounted for by the fall having occurred during the winter, when there was a very heavy cover of snow and ice, which might not have been entirely dissipated by the impact. The possibility that the mass may have fallen elsewhere during glacial time, and been transported by the ice-sheet, seems rather far-fetched, especially

since no glacial striae are visible, and the weathered crust is relatively thin. While it is by no means a universal rule, meteorites that have been found in glacial drifts usually have a much thicker crust than this mass.

Nomenclature

This mass will be known as the Osseo Iron, very coarse kamacitic octahedrite.

Acknowledgements

Thanks are due to the Assistant Secretary of the United States National Museum for permission to use the laboratory and other facilities of the Museum for this investigation. The kind assistance of Dr. W. F. Foshag and Mr. E. P. Henderson in the description of the mass and the choice of analytical methods are most gratefully acknowledged.

ZEOLITES IN NEW MEXICO

C. E. Needham, *School of Mines, Socorro, New Mexico.*

The writer has been unable to find any occurrences of zeolites described from New Mexico. Consequently, the two localities mentioned below are possibly the first on record from this state. Both localities are near Socorro, the first about twelve miles northeast, the other about three miles west of Socorro.

Valle del Ojo de la Parida

The Valle del Ojo de la Parida is a depression lying in the northeastern corner of the Socorro quadrangle. It represents a remnant of a great area of pediment[1] cut mainly in Paleozoic, Mesozoic, and Tertiary rocks. Residual hills over the valley are capped with the Santa Fe formation of late Pliocene age, the main body of the Santa Fe being largely eroded from this immediate vicinity. Bordering the valley on the west is a belt of Tertiary volcanic rocks. It is in one of these volcanic flows that the zeolites are found.

The zeolite rock is a dark colored andesite, possibly extruded in Santa Fe time and covered shortly after extrusion with Sante Fe sediments. Although several andesite flows are found within the Santa Fe formation near Socorro, only this one is known to contain zeolites.

The Sante Fe formation in this locality is a very coarse, poorly sorted gravel. It is clearly of fluvial origin, swept out from the nearby hills by ephemeral streams and deposited as alluvial fans along the fronts of the

[1] Bryan, Kirk, The formation of pediments: *Rpt. of XVI International Geological Congress, Washington.* Preprint, 1933.

hills. Subsequent erosion has stripped away the Santa Fe sediments, leaving the andesite uncovered and weathered.

Thomsonite. The most abundant zeolite in the andesite is thomsonite. It is white and has a radial structure. It shows perfect cleavage parallel to [010], fuses easily to a white enamel, and gelatinizes with hydrochloric acid. Its optical properties are as follows: weak birefringence, biaxial, optically positive, $2V=55°$ (estimated), $\gamma=c=$elongation, α near 1.517, β near 1.519, γ near 1.528. It fills cavities and lines cracks in the altered andesite.

Analcite. Associated in very small amounts with the thomsonite is a colorless mineral occurring in irregular glassy grains. Most of the grains are isotropic and have an index of refraction near 1.490. Other grains show very weak birefringence and exhibit cross twinning to a marked degree. This mineral evidently is analcite.

Natrolite. Also in small amounts with the thomsonite and analcite is colorless to white natrolite. Its optical properties are: biaxial positive, weak birefringence, $2V=60°$ (estimated), $\beta=1.482\pm.001$.

Socorro Mountains

The Socorro Range rises about three miles west of Socorro and extends approximately parallel to the Rio Grande for several miles to the north and south. It represents great piles of Tertiary flows and pyroclastic rocks resting on Pennsylvanian sediments. The range lies along a zone of strong faulting, most of the faults striking roughly with the elongation of the range. Much of the faulting took place in post-Santa Fe time, that is, at the close of the Pliocene, or at the beginning of the Pleistocene. The faulting was accompanied, or followed shortly, by ascending thermal waters, the alteration effects of which are plainly seen in a number of places along the front of the range.

Near the base of the Tertiary sequence are beds of andesitic breccia and white tuffs. In places the tuffs grade upward and laterally into rhyolitic material, and in other places into a coarse sand and gravel consisting almost entirely of fragments of vitrophyre. The glass of the vitrophyre has an index of refraction of $1.487\pm.002$ and thus is seen to be highly silicic. In other exposures fragments of andesite and quartz are common in the tuffs, also some glass with an index of refraction of $1.507\pm.003$.

Both breccia and tuff have undergone considerable hydrothermal alteration along the fault zone. The most common alteration product in them is a cream-colored to pale green clay mineral, identified as beidellite. Along the north side of Blue Canyon the bentonitic tuff contains quantities of a zeolite whose optical properties are identical with the

thomsonite described from Valle del Ojo de la Parida, except β is near 1.516.

Here, however, the thomsonite occurs as irregular clusters of botryoidal forms imbedded in the altered tuff. The mineral is commonly pale pink and has a radial structure. Thus far, no other zeolite has been found in the tuff.

Origin of the Zeolites

Two theories may be suggested to explain the origin of the zeolites in the andesite in Valle del Ojo de la Parida. The first is that they were formed by the interaction between the dissolution products of the andesite and the meteoric waters percolating through the Santa Fe gravels. The second is that the zeolites resulted from the interaction of hot solutions from the lava with certain early formed silicate minerals in the andesite.

It is doubtful if evidence is at hand to prove either theory. The writer believes that mineralized magmatic waters are more effective than meteoric waters in altering silicates to zeolites.

Regarding the thomsonite in the Socorro Range, the writer believes the mineral was formed by the hydrothermal alteration of a tuff consisting essentially of siliceous volcanic material admixed with some quantity of an intermediate composition. The alteration of siliceous rocks to zeolites is undoubtedly a much less common phenomenon than the alteration of basic rocks to the same minerals.

THE POCKET MINERALOGICAL OR POLARIZING MAGNIFIER

V. V. Arshinov, *Institute of Economic Mineralogy, Moscow, U.S.S.R.*

The author had the pleasure of demonstrating both to the members at the meeting of mineralogists of U.S.S.R., held in Moscow in May 1937, and to the members of the XVII Session of the International Geological Congress, a small instrument, shown in figure 1. This instrument, named "the pocket mineralogical or polarizing magnifier," permits convenient identification in the field of non-opaque minerals and rocks by petrographic-optical methods. The instrument consists of a disc, graduated on the circumference, holding the object glass and attached to a handle. On the two sides of the disc round polaroid plates are placed, which may be rotated simultaneously, with reading of the degrees of rotation on the circumference of the disc. Over the polaroid-polarizer, directly under the object glass, a round glass plate is placed, rotating simultaneously with the polarizer, with a cross cut on it.

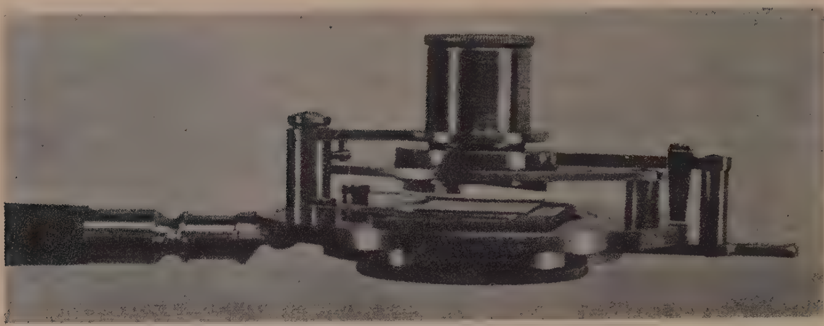


FIG. 1

Below the glass plate is a slit for inserting the compensators (gypsum plate, red 1st order, or graduated quartz wedge) and light filters.

Placed over the polaroid-analyzer there is the magnifier (×10 or ×16). Divisions on the quartz wedge may be seen through the lens simultaneously with the outlines of grains or of sections of minerals in slides of the rock, and the path difference of birefringent minerals may be determined by the compensation band.

In order to observe pleochroism, the polaroid-analyzer of the polaroid-polarizer may be removed from the ring in which they rotate. To compensate the slight coloration of the polaroid, a plate with a corresponding light filter may be inserted in the slit over the polarizer or in place of the polarizer or analyzer.

For observations of conoscopic interference figures, e.g., of a muscovite plate, the magnifier is swung to one side and the phenomenon is observed by placing the polaroid-polarizer or the polaroid-analyzer close to the eye.

In order to observe the interference figures in a section of a small grain of a mineral in a rock slide, preferably not covered by a glass, the microlens of the microconoscope is introduced over the mineral grain. The preparation of the microconoscope was described by N. Vedeneeva and A. Kolotushkin.[1] Accurate microballs of a small size, about 1 mm. in diameter are also readily obtained by another simple method described in the Lomonosov Institute of the Academy of Sciences. Instead of a microball, half immersed in a celluloid plate, use may be made also of a cover glass with a hemispherical lens (radius of curvature 0.5–0.6 mm.) cemented to it with Canada balsam. A thin glass plate with some air bubbles also can be used as a microconoscopic plate.

The disc with polaroids may be inserted in a slit of a block of wood or ruler (24 cm. long, 3–4 cm. high and 2–3 cm. wide). Since any diam-

eter of the disc may be placed parallel to the length of the ruler, the combination of the disc with the polaroids and a ruler may replace to a certain extent the conoscope and the Fedorov stage with two axes of rotation. The angle of the optic axes in a mica plate may be determined by means of the instrument in the following way. The ruler with the disc (on the level of the eyes arranged by means of a photographic stand) is placed upon a sheet of paper, fastened with knobs on the plate trestle board. By moving the ruler with the attached disc on the sheet of paper, the observed point of emergence of each of the optical axes of a mica plate is aligned with some signal, e.g., a telegraph pole, or a straight trunk of a tree which is seen through the mica and polaroids, and corresponding lines are drawn on the paper with the aid of the ruler.

The picture of the optic axes may be fixed also at the cross of the frosted glass of the photographic apparatus.

The angle between both the lines that have been drawn on the sheet of paper will correspond to the angle of the optic axes of the mica.

To the same end the Wulff net may be used by rotating the ruler about the center of the net and reading the angles on the circumference of the net.

For the observations of the phenomena taking place in characteristic directions of small mineral grains (<3 mm. >0.3 mm.) in thin sections of rocks, and for measuring the angles between these directions, the disc with polaroids in combination with a telemagnifier or a telemicroscope, like those made, for instance, by Zeiss,[2] may be used.

The maintenance of extinction while rotating the ruler with the disc around the axis perpendicular to the ruler and parallel to one of the two directions of extinction in the section of the mineral, will be evidence that the latter is uniaxial.

A disc of greater radius with an attached mechanism for a parallel shifting of the object glass with a rock slide may be used in petrofabric investigations, for measuring the angles between the optic axis and the perpendicular to the slide of such minerals as quartz and calcite. In the case when the optic axis makes a small angle with the plane of the thin section a method is applied like that described in the paper "On glass hemispheres for the petrographical microscope."[3] The investigations in the Petrographic Laboratory of the Institute of Economic Mineralogy have shown that the determination of the refractive indices by the immersion method with a double diaphragm and oblique illumination, may be made with this instrument with the magnifier $\times 10$, with an accuracy within five units in the third decimal place.

By placing the pocket mineralogical or polarizing magnifier in front of the microscope lamp, or a classroom lantern, good pictures of rock

microsections in beautiful interference colors may be obtained on a screen, using the magnifier of the instrument as an objective.

References

1. Vedeneeva, N., and Kolotushkin, A., The conoscope with modified bulb: *Trans. Institute Economic Mineralogy, Moscow, U.S.S.R.*, no. **54,** pp. 32–42.
2. Zeiss, Fernrohrlupen und Fernrohrmikroskope, Med. 3, (7), 1936.
3. Arshinov, V. V., *Trans. Institute Economic Mineralogy, Moscow, U.S.S.R.*, no. **65,** p. 42, 1934.

PROCEEDINGS OF SOCIETIES

NEW YORK MINERALOGICAL CLUB

American Museum of Natural History, New York City, February 16, 1938

The meeting was called to order by first Vice-President Olaf Andersen at 8.15 P.M. with 135 members and guests present. Professor S. J. Shand of Columbia University addressed the Club upon "Mineral Deposits of South Africa." Dr. Shand began with a summary of the general geology of South Africa and the neighboring states and indicated in general the relation of the deposits to the various formations. He then spoke upon specific deposits and gave interesting details of occurrences and associations which are not generally known. The Rand quartzites were described as containing, among other minerals, zircon, chromite, platinum, and osmiridium; and most interesting, small green diamonds, as primary minerals. He spoke of pyrite, chloritoid, muscovite, and tourmaline among the secondary minerals and then gave the data on the gold occurrence with the evidence relating to the controversy as to its origin. One very interesting specimen showed a streak of rounded pellets of carbon with a concentration of minute but visible gold flakes in the immediate vicinity.

Platinum was next discussed including two types of occurrences in the Bushveld, both in the norite and in dunite pipes cutting the latter and carrying up to 8 oz. a ton in their richest portions. Chromite and tin were also discussed, and a very unusual occurrence of the latter described. This was in the northern part of the Complex, where surface exposures of granite showed rings of rock impregnated with small black tourmaline needles and masses up to 20 feet in diameter. Within these rings were rich aggregates of cassiterite and mica, running as high as 80% cassiterite. Two of these pipes, most of which are now mined out, extended as much as 700 feet in depth.

Lead, zinc, manganese, copper, iron, coal and oil deposits were referred to briefly, and a new occurrence of zunyite described. Diamonds and their history received only a brief mention since they are so well known, but the members were interested to hear that the grease tables are so effective in retaining all of the diamonds which flow over them that it is rare to find a crystal farther down than the first ten inches of the table.

The crocidolite asbestos deposits were described but their origin was not explained. The "Tiger's eye" semi-precious gem material develops naturally from the fibrous crocidolite through oxidation and leaching of the other elements, leaving the silica to form a pseudomorph of the asbestos fibers. Other types of asbestos and the South African occurrence of stichtite were briefly mentioned and a fine specimen of this attractive mineral displayed.

Pegmatites are abundant in South Africa and show many of the characteristics of the American pegmatites. In the eastern Transvaal, pegmatites cut granite and basic schists,

with interesting variations in their composition. In the granite, the pegmatites are rich in muscovite, and one crystal was mined which measured 16 by 10 feet. The pegmatites in the basic schists, on the other hand, are deficient in silica and fine large corundum crystals occur in the feldspar. Several have been collected that measured two feet in length, though most of the weathered-out crystals are now gone. Some show fine color and Dr. Shand exhibited two fine red translucent specimens, but said thus far nothing of gem quality has been obtained. Along the margins of some of these dikes, in the schist, good emerald crystals have been mined.

Complex pegmatites are also to be seen in Namaqualand which contain beryl, chrysoberyl and associated lithium minerals, but none of these minerals are present in commercial amounts. Later Dr. Shand added a few remarks on some interesting rocks, especially soda-rich rocks some containing remarkably large nepheline crystals. A manganiferous pectolite and rich eudialyte-bearing rocks were shown; rocks that are similar to those of Magnet Cove. The talk was well illustrated by specimens and maps and was very informative because of the many sidelights which were the results of personal experiences in the field.

At the conclusion of Prof. Shand's lecture, L. N. Yedlin showed some polished specimens of Franklin minerals.

Meeting of February 18, 1938

The Club held a special meeting in the Auditorium of The American Museum of Natural History on the evening of Feb. 18th to hear an address by Mr. H. S. Spence of the Canadian Bureau of Mines and Resources. This talk was sponsored by the club, the American Museum and the New York Academy of Sciences. Nearly one thousand members of the combined organizations and their guests were present. Mr. Spence spoke upon "Mining Radium in Canada's Subarctic" and illustrated it with specimens, slides and moving pictures. The display included several fine specimens of pitchblende and native silver, and included one of the first specimens found by Labine. The slides and moving pictures showed the different settlements passed on the trip to Great Bear Lake and the development of the property to its present state. At the end he spoke for a few minutes upon the recovery of radium from pitchblende and showed pictures of the various stages in the process.

F. H. Pough, *Secretary*

PHILADELPHIA MINERALOGICAL SOCIETY

Academy of Natural Sciences of Philadelphia, December 2, 1937

A stated meeting was held with Mr. Trudell in the chair, and 47 members and 28 visitors in attendance.

Dr. Harry H. Hess addressed the society on "Mineral Collecting in Russia" descriptive of trips to the Kola Peninsula and the Ural Mountains during the International Geological Congress in Moscow. Localities visited included the deposit of metamorphosed coal on Lake Onegin, and feldspar quarries on islands in the lake; the Chibine complex where deposits of apatite and lovchorrite (a titanosilicate of cerium) in the alkaline syenites are being exploited; and deposits of chromite, magnetite, platinum, gold, nickel, and asbestos in the Urals. Specimens were exhibited.

Dr. Berliner exhibited flexible sandstone (locally called "limber-grit") from Rural Hall, North Carolina. Mr. Morgan reported finding beraunite at Hellertown, Pa.

Louis Moyd, *Secretary*

Academy of Natural Sciences of Philadelphia, January 6, 1938

A stated meeting was held with Mr. Trudell in the chair, and 43 visitors and 26 members present.

Mr. Adolph Meier and Mr. Harold Tomlinson described the occurrence of harmotone and montmorillonite near Glen Riddle, Delaware County, Pa. Dr. Edgar T. Wherry presented a résumé of the annual meetings of the Geological Society and the Mineralogical Society held in Washington.

Academy of Natural Sciences of Philadelphia, February 3, 1938

A stated meeting of the Philadelphia Mineralogical Society was held on the above date with the president, Mr. Trudell, in the chair; 53 members and 25 visitors were present.

Mr. Arthur Montgomery addressed the society on Uta Phosphates and Montana Amethysts." During the summer of 1937, Mr. Montgomery, working with Mr. Edwin Over, reopened the old variscite mines at the edge of the Great Salt Lake desert, in Fairfield Co., Utah.

The phosphate minerals occur as nodules replacing limestone along a steeply pitching fracture zone. The earliest mineral of the phosphate series is variscite, which may be replaced by wardite, gordonite, and finally by pseudowavellite. Owing to the inaccessibility of the locality, some difficulty was experienced in transporting mine timbers and other equipment.

Later in the season, Messrs. Montgomery and Over worked a pegmatite about 25 miles S. E. of Butte, Montana. Large vugs in the dike, some over fifteen feet in diameter, were lined with amethyst, smoky quartz, tourmalinated quartz and feldspar crystals. In closing, the speaker said that because of the past season's success, he would continue his work of developing localities containing unusual mineral species. Mr. Montgomery illustrated his lecture with many beautiful polished sections of the phosphate nodules and with crystals from the Montana locality.

Mr. John Boyle reported on the finding of quartz crystals and etched quartz at the Gaylor Quarry in Saratoga Springs, N. Y. Mr. Stephen Varni exhibited a 264 carat cut emerald from the Urals.

LOUIS MOYD, *Secretary*

NEW MINERAL NAMES

Aminoffite

CORNELIUS S. HURLBUT: Aminoffite, a new mineral from Långban. *Geol. Fören. Förhandl. Stockholm*, vol. **59**, No. 410, pp. 290–292, 1937.

NAME: In honor of Dr. G. Aminoff.

CHEMICAL PROPERTIES: A hydrous silicate of calcium, beryllium and aluminum: $Ca_{24}Be_9Al_3Si_{24}O_{84}(OH)_3 \cdot 12H_2O$. Analysis (by F. A. Gonyer) SiO_2 42.49, Al_2O_3 4.41, BeO 6.20, Fe_2O_3 0.31, MnO 0.19, CaO 40.27, H_2O 6.45; Sum 100.33. Insoluble in acids. Infusible.

CRYSTALLOGRAPHICAL PROPERTIES: Tetragonal $c=0.7116$. $(111).(001)=45°11'$. Forms p (111) and c (001). Habit simple pyramidal. $a=13.8$ Å. $c=9.8$ Å. Space group $I\,4/m\,m\,m$.

PHYSICAL AND OPTICAL PROPERTIES: Clear and colorless; luster vitreous. Uniaxial, sometimes anomalously biaxial. Negative. $\omega=1.647$, $\epsilon=1.637$. Brittle. Cleavage (001) poor. Fracture conchoidal. Hd. 5.5. G=2.94.

OCCURRENCE: Found as small crystals in veins or cavities in massive magnetite or limonite.

RELATIONSHIPS: Related to meliphanite $(Ca, Na)_2Be(Si, Al)_2(O, F)_7$.

W. F. FOSHAG

Scheteligite

HAROLD BJØRLYKKE: Scheteligite, a new mineral. *Norsk. Geol. Tids.*, vol. **17**, pp. 47–49, 1937.

NAME: In honor of Prof. Jacob Schetelig, late director of the Mineralogical Museum in Oslo.

CHEMICAL PROPERTIES: A titanate, tantalate, columbate and tungstate of calcium, manganese, etc.; $(Ca,Fe,Mn,Sb,Bi,Y)_2 \cdot (Ti,Ta,Nb,W)_2(O,OH)_7$. Analysis (by V. Bruun): CaO 10.73, FeO 1.88, MnO 6.19, Sb_2O_3 7.77, Bi_2O_3 2.54, Y_2O_3 6.00, TiO_2 18.73, Ta_2O_5 20.00, WO_3 5.00, Nb_2O_5 8.65; loss on ignition 2.00, SiO_2, calculated as microline, 9.70; Sum 99.19. Soluble in hydrofluoric acid.

CRYSTALLOGRAPHICAL PROPERTIES: Orthorhombic, with rough pyramid faces.

PHYSICAL PROPERTIES: Color black. Streak pale yellow to grayish; in thin section reddish brown. Fracture conchoidal. Hd. 5.5. G. 4.74.

OCCURRENCE: Found as small embedded crystals in a pegmatite at Torvelona, Iveland, Norway, together with tourmaline, bismuth, alvite, euxenite, monazite, thortveitite, beryl, etc.

W. F. F.

Magallanite

G. A. FESTER, J. CRUELLAS AND F. GARGATAGLI: La Magallanita, un nuevo Mineral bitumenoso. *Anal. Soc. Cient. Argentina*, vol. **124**, p. 211, 1937. An asphaltic substance found as rolled pebbles near Magallanes. No analysis or other data given.

W. F. F.

Gahno-spinel

B. W. ANDERSON AND C. J. PAYNE: Magnesium-zinc-spinels from Ceylon. *Mineralog. Mag.*, vol. **24**, No. 158, pp. 547–554, 1937.

Some pale blue to deep blue gem spinels from Ceylon have indices of refraction of 1.7153–1.7469 and specific gravities of 3.584–3.981. They are intermediate between spinel and gahnite and contain zinc. The name "gahno-spinel" has been proposed.

W. F. F.

β-Ascharite

M. N. Godlevsky: Mineralogical investigation of the Inder borate deposit. *Mem. Soc. Russ. Mineral.*, 2d *Ser.*, vol. **66,** No. 2, pp. 315–344, 1937, 11 text figures, 5 plates, 1 map (in Russian); pages 345–368 in English.

Chemical Properties: A hydrous magnesium borate: $MgHBO_3$. Analysis (E. N. Egorova, analyst): SiO_2 0.20; Al_2O_3 0.16; Fe_2O_3 0.13; MnO 0.02; CaO none; MgO 46.70; Na_2O 0.20; K_2O 0.26; B_2O_3 40.85; Cl 0.11; SO_3 0.69; CO_2 None; H_2O+ 10.95; H_2O- 0.32. Sum 100.59. Insoluble in water, slowly soluble in acids. B.B. fuses to an enamel.

Crystallographical Properties: Orthorhombic?

Physical and Optical Properties: Color white, luster chalky. Biaxial, 2 V small. $\alpha=1.575$, $\beta=1.642$, $\gamma=1.646$. Parallel extinction; X parallel to elongation of the fibers. Hd. 3.5. G. 2.65.

Occurrence: The principal industrial mineral in the Lake Inder borate deposits, near the Ural River, about 150 km. north of the Caspian Sea. Found as chalky masses of fine fibrous habit or massive. Differs from the ascharite of Germany (here called *α-ascharite*) in optical properties. Closely related to camsellite.

W. F. F.

Inderite

Ibid.

Chemical Composition: A hydrous magnesium borate: $Mg_2B_6O_{11}\cdot 15H_2O$. Analysis (by E. N. Egorova): SiO_2 0.13; Al_2O_3 0.02; Fe_2O_3 0.33; MnO none; CaO 0.16; MgO 14.65; Na_2O 0.17; B_2O_3 36.20; CO_2 0.17; H_2O+ and H_2O- 48.20; Sum 100.03.

Physical and Optical Properties: Color white, luster vitreous. Biaxial, 2 V large. $\beta=1.488$, $\gamma=1.504$, $Z\wedge c=5°$. G=1.80.

Occurrence: Found as small nodules or aggregates of minute needles.

W. F. F.

Gedroitsite

I. N. Antipov-Karatajev and I. D. Sedleckij: On the genesis of colloidal metals in salt soils. Gedroitsite, a new mineral. *Compt. Rend. Acad. Sci., U.S.S.R., new ser.* vol. **17,** No. 5, pp. 251–254, 1937.

A name proposed for an artificial crystalline aggregate produced by adding sodium aluminate solution to sodium silicate solution and allowing the resultant precipitate to stand for several years. The final product is chiefly crystalline, isotropic, with $n=1.483$.

W. F. F.

Correction

In the editor's report for 1937 (*Am. Mineral.*, vol. **23,** p. 162) the statement is made that the current volume describes six new minerals. The correct number is seven. The new mineral species "ellestadite" was inadvertently omitted and should be added to the list.